UNIVERSITÉ DU QUÉBEC À MONTRÉAL

CARBON ACCUMULATION IN THREE OMBROTROPHIC PEATLANDS OF THE EASTMAIN REGION, QUEBEC, CANADA: INFLUENCE OF HYDROLOGY, VEGETATION AND FIRE IN THE HOLOCENE CLIMATIC CONTEXT

DISSERTATION
PRESENTED
AS PARTIAL REQUIREMENT
OF THE DOCTORATE IN ENVIRONMENTAL SCIENCES

BY
SIMON VAN BELLEN

JUNE 2011

UNIVERSITÉ DU QUÉBEC À MONTRÉAL

ACCUMULATION DU CARBONE DANS TROIS TOURBIÈRES OMBROTROPHES DE LA RÉGION D'EASTMAIN, QUÉBEC, CANADA : INFLUENCE DE L'HYDROLOGIE, DE LA VÉGÉTATION ET DES FEUX DANS LE CONTEXTE CLIMATIQUE HOLOCÈNE

THÈSE
PRÉSENTÉE
COMME EXIGENCE PARTIELLE
DU DOCTORAT EN SCIENCES DE L'ENVIRONNEMENT

PAR
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JUIN 2011

AVANT-PROPOS

Cette thèse est composée de trois articles qui forment chacun un chapitre et qui ont été soumis ou acceptés dans des revues scientifiques avec comité de lecture. Pour chaque article, je suis le principal responsable de la récolte des données incluant l'échantillonnage au terrain, les analyses de laboratoire, l'interprétation des résultats et la rédaction. Les co-auteurs des articles ont contribué à la recherche par des échanges sur les résultats et interprétation de la recherche, l'aide à l'utilisation des logiciels ou encore en fournissant des données.

Le premier article s'intitule «Quantifying spatial and temporal Holocene carbon accumulation in ombrotrophic peatlands of the Eastmain region, Quebec, Canada» et a été publié dans la revue *Global Biogeochemical Cycles* en mai 2011. Les co-auteurs de cet article sont Pierre-Luc Dallaire (ex-candidat à la maîtrise), Michelle Garneau et Yves Bergeron (respectivement directrice et codirecteur de recherche). Dans ce chapitre, les données issues de l'étude du terrain à l'aide du géoradar ont été fournies par Pierre-Luc Dallaire.

Le deuxième article est sous presse dans la revue *The Holocene* et s'intitule «Holocene carbon accumulation rates from three ombrotrophic peatlands in boreal Quebec, Canada: impact of climate-driven ecohydrological change». Il est co-écrit par Michelle Garneau et Robert K. Booth. La contribution de Robert K. Booth (Lehigh University, États-Unis) a consisté à l'application d'une fonction transfert appliquée aux assemblages de thécamoebiens et permettant d'obtenir une reconstruction de la nappe phréatique dans le temps.

Le troisième article, «Did fires drive Holocene carbon sequestration in boreal ombrotrophic peatlands of eastern Canada?» a été soumis pour publication à la revue *Quaternary Research* en décembre 2010. Les co-auteurs sont Michelle Garneau, Yves Bergeron et Adam A. Ali (Université Montpellier II). Adam A. Ali a contribué à l'application du logiciel utilisé à l'interprétation des séquences de macrocharbons.

REMERCIEMENTS

Un grand merci à ma directrice de thèse, Michelle Garneau, pour m'avoir donné l'opportunité de participer à ce magnifique projet, et pour les discussions et le soutien qui ont été essentiels à la réalisation de la thèse. Je remercie Yves Bergeron, pour sa contribution à la compréhension de la dynamique des feux. Merci à Dr. Alain Tremblay (conseiller scientifique senior du projet GES – Eastmain-1, Hydro-Québec) pour son support financier et logistique dans le cadre du projet Eastmain-1. Merci également au Fonds Québécois de la Recherche sur la Nature et les Technologies (FQRNT : fonds forestiers, subvention à Yves Bergeron) pour le financement qui m'a été octroyé.

La recherche n'aurait pas pu être réalisée sans la contribution de Hans Asnong aux campagnes de terrain ainsi qu'à la gestion du laboratoire et des données. Un grand merci à Pierre-Luc Dallaire pour son assistance au terrain et à la cartographie des sites. Merci à Luc Pelletier pour le travail de logistique lors des campagnes de terrain. Je remercie également plusieurs assistant(e)s dont la contribution aux campagnes de terrain et au laboratoire a été très importante, notamment Robin Beauséjour, Maxime Boivin, Diane Charbonneau, Noémie Cliche Trudeau, Marilou Hayes, Géraldine St-Pierre, Anny Tadros, Benoît Talbot et Antoine Thibault.

Je remercie également Dr. Adam A. Ali (Université Montpellier 2, France), Dr. Hugo Asselin (Université du Québec en Abitibi-Témiscamingue), Dr. Robert K. Booth (Lehigh University, États-Unis) et Bertrand Fournier (Université du Québec à Montréal) pour leurs conseils et commentaires sur les articles.

Un remerciement au groupe Les Tourbeux, et spécialement à Pierre-Luc Dallaire, Aurélie Genries et Anne Quillet, pour l'inspiration, les discussions ainsi que la motivation et les encouragements. Merci finalement à Marie-Ève Ménard pour la correction du français.

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LIST OF ABBREVIATIONS, ACRONYMS AND SYMBOLS

^{14}C	Carbon-14 isotope
AGC	Automatic Gain Control
AMS	Accelerator Mass Spectrometry
C	Carbon
cal BP	Calibrated years before present (AD 1950)
CH_4	Methane
CHAR	Charcoal Accumulation Rate
CO_2	Carbon dioxide
DECLIQUE	Dynamique des Écosystèmes et Changements Climatiques Chair
DGPS	Differential Global Positioning System
EM-1	Eastmain-1
F-fire	Forest fire
FP-fire	Forest/Peatland fire
GOF	Goodness-Of-Fit
GPR	Ground-Penetrating Radar
IPCC	Intergovernmental Panel on Climate Change
KOH	Potassium hydroxide
LIA	Little Ice Age
LLC	Lac Le Caron
LOI	Loss-On-Ignition
LORCA	Long-term Apparent Rate of Carbon Accumulation

LOWESS	Locally Weighted Scatterplot Smoothing
MAP	Mean Annual Precipitation
MAT	Mean Annual Temperature
MOS	Mosaik
MWP	Medieval Warm Period
NEP	Net Ecosystem Productivity
NLWIS	National Land and Water Information Service
NPP	Net Primary Production
OM	Organic Matter
PAR	Photosynthetically Active Radiation
PVC	Polyvinyl chloride
RMSEP	Moot-Mean-Square Error of Prediction
SE	Standard Error
SNI	Signal-to-Noise Index
spp.	Species (plural)
STE	Sterne
UOM	Unidentifiable Organic Matter
UQAM	Université du Québec à Montréal
WTD	Water Table Depth

RÉSUMÉ

Les tourbières nordiques accumulent du carbone (C) puisque dans ces milieux, la production de la matière organique excède sa décomposition, ce qui a résulté en un important réservoir de C représentant environ 30% du C présent dans les sols terrestres à l'échelle de la planète. Puisque l'accumulation du C durant l'Holocène a influencé le climat global, les projections des changements climatiques devraient tenir compte de cette dynamique du C dans ces écosystèmes. Des variations de l'accumulation du C sont généralement liées à une combinaison de facteurs géomorphologiques (topographic du bassin), climatiques, des processus écologiques autogènes et des évènements ponctuels de perturbation.

À part les feux, qui représentent une émission directe de C vers l'atmosphère, des facteurs internes et externes influencent la dynamique du C dans les tourbières. Les variations hydrologiques constituent un facteur déterminant en ce qui a trait aux assemblages végétaux. De plus, la végétation influence l'hydrologie par les processus de transpiration et d'isolation de la tourbe. L'objectif principal de la thèse était de quantifier les stocks de C ainsi que les taux d'accumulation pour trois tourbières boréales situées dans le nord du Québec. Nous avons de plus estimé l'influence des changements de la végétation, de l'hydrologie et de l'intensité et de la fréquence des feux de tourbière sur les taux d'accumulation du C, en tenant compte du contexte climatique holocène du nord du Québec.

Afin de quantifier le volume de tourbe de chaque tourbière, les profondeurs ont été mesurées par sondage manuel ainsi que par un géoradar (GPR). La stratigraphie de plusieurs carottes échantillonnées dans chaque tourbière a été analysée. L'interpolation spatiale des taux d'accumulation du C a permis une reconstitution de chaque écosystème. Les changements de la végétation et de la nappe phréatique durant l'Holocène ont été reconstruits à l'aide d'analyses de macrorestes végétaux et de thécamoebiens. Les régimes de feu ont été reconstruits à partir du dénombrement de charbons de bois macroscopiques. Les variations temporelles des assemblages végétaux, de la nappe phréatique et de l'intensité et de la fréquence des feux ont été comparées aux fluctuations de l'accumulation du C.

Les trois tourbières étudiées ont accumulé du C à un taux moyen de 16,2 g m^{-2} an^{-1} depuis l'âge maximum de 7510 cal BP, ce qui équivaut à une masse moyenne de C de 91 kg m^{-2} et un réservoir total de C de 608×10^6 kg. Au début de leur développement, l'expansion latérale des tourbières a été rapide, tandis que l'accumulation du C à l'échelle de l'écosystème a culminé entre 5250 et 3500 cal BP. Malgré le fait que les taux d'accumulation verticale aient été généralement élevés dès le début du développement des tourbières, la topographie des bassins dans lesquels se sont développés ces écosystèmes a limité l'accumulation générale de la tourbe.

Les résultats montrent que dans l'ensemble, les assemblages dominants de végétation ont varié dans le temps ainsi qu'entre chaque tourbière. Les périodes où les taux d'accumulation ont été élevés dans les tourbières de LLC et STE étaient dominées par une végétation de *Sphagnum* section *Acutifolia*, ainsi que par des nappes phréatiques intermédiaires entre 10-15 cm en dessous de la surface. Le ralentissement de l'accumulation du C durant l'Holocène récent a été associé aux fluctuations importantes de la nappe phréatique ainsi qu'à une diminution des sphaignes de la section *Acutifolia*. En effet, la présence de ces sphaignes limite les taux de décomposition en maintenant des conditions froides, acides, humides et faibles en nutriments, favorisant l'accumulation du C dans ces

tourbières. En général, les tourbières MOS et STE montrent une moins grande présence de sphaignes dans leurs sections centrales, probablement dû au contexte physiographique local des bassins dans lesquelles elles se sont développées en maintenant des conditions minérotrophes jusqu'à 5450 et 4410 cal BP, respectivement.

Les analyses de macrocharbons contenus dans la tourbe ont permis d'identifier des intervalles moyens de feu de 883 ans durant l'Holocène, même si d'importantes variations spatiales et temporelles ont été reconstituées. La fréquence des feux et la production de charbons a augmenté après 2000 cal BP. Néanmoins, les analyses de régression montrent que les feux n'ont pas été un facteur déterminant dans l'accumulation du C. De plus, les changements du régime des feux de l'Holocène récent semblent avoir été indépendants des changements dans la végétation locale.

Le refroidissement néoglaciaire de ~3000 cal BP pourrait bien avoir été le facteur principal ayant engendré ces changements environnementaux. Ce refroidissement climatique a probablement affecté de façon négative les taux d'accumulation du C. De plus, les périodes synchrones de nappe phréatique basse durant l'Holocène récent ont probablement été une des principales causes du changement de la végétation, dominée alors par des espèces ligneuses aux dépens des sphaignes, une tendance observée notamment à LLC.

Afin d'obtenir un portrait précis de l'accumulation à long terme du C, les reconstructions futures devraient tenir compte des variations de l'expansion latérale des tourbières, puisque les masses de C et les taux de l'accumulation issus de carottes centrales surestiment les taux à l'échelle de l'écosystème. Les changements climatiques futurs pourraient impliquer des augmentations à la fois de la nappe phréatique et des températures, ce qui aurait probablement un effet positif sur la croissance des sphaignes, et donc sur l'accumulation du C dans les tourbières de la région.

Mots clés: tourbière, carbone, Holocène, feu, paléoécologie

ABSTRACT

Northern peatlands sequester organic carbon (C) due to a long-term inbalance between primary production and decomposition of organic matter, which has resulted in a globally important C stock, representing ~30% of global terrestrial soil C. As Holocene peatland C sequestration has influenced global climates, climate projections should take account of C dynamics in these ecosystems. Variations in net C sequestration are generally driven by a combination of geomorphology (e.g. basin topography), climate, disturbance (e.g. fire) and autogenic processes.

Except fire, which represents a direct emission of C to the atmosphere, autogenic and allogenic factors primarily influence C sequestration through ecosystem hydrology, while hydrology itself is a principal driving factor for local vegetation assemblages. Finally, vegetation feeds back on hydrology by affecting transpiration and temperature of the underlying peat. We aimed to quantify total C stocks and C accumulation rates at the ecosystem scale in three pristine boreal peatlands in northern Quebec. In addition, we estimated the effect of changes in vegetation, hydrology and peatland fire regimes on C accumulation rates, taking into account the Holocene climatic context.

Peatland depth was measured using manual probing and Ground-Penetrating Radar analyses. Multiple cores from each peatland were sampled and stratigraphically analyzed. Spatial interpolation of C accumulation patterns resulted in a reconstruction at the scale of each ecosystem. Holocene vegetation and water table heights were reconstructed by plant macrofossil and testate amoeba analyses and fire intensity and frequency by macroscopic charcoal, and trends were compared to changes in accumulation.

The three studied peatlands have sequestered C at an average rate of 16.2 g m^{-2} yr^{-1} since a maximum age of 7510 cal BP, resulting in a area-weighted mean C mass of 91 kg m^{-2} and a total C stock of 608 × 10^6 kg. Holocene patterns show important rates of lateral expansion in the early development stages of each ecosystem, while ecosystem-scale rates of C accumulation culminated between 5250 and 3500 cal BP. Although vertical accumulation rates were high shortly after peatland initiation, the limited areal extent of each ecosystem initially prevented an important C sink functioning.

Dominant vegetation assemblages varied among cores and during the Holocene. Mid-Holocene episodes of important vertical accumulation in LLC and STE bog coindiced with dominant *Sphagnum* section *Acutifolia* cover and intermediate water table depths of around 10-15 cm below the surface, while the late-Holocene slowdown was associated with highly fluctuating water tables and decreases in the presence of *Sphagnum* section *Acutifolia*. These *Sphagna* are likely a determinant factor in restricting decomposition rates by creating of cold, acidic and nutrient-poor conditions, and therefore positively influence C sequestration. MOS and STE peatlands generally show a less important *Sphagnum* cover in their central sections, probably due to the local geomorphological context of their basins, causing minerotrophic conditions to persist until 5450 and 4410 cal BP, respectively.

Analyses of macroscopic charcoal resulted in a quantification of Holocene mean peatland fire intervals of 883 years, yet important spatial and temporal variability was registered. Peatland fire frequency and charcoal production generally increased after 2000 cal BP. However, regression analyses indicate that fires have not been a driving factor

for C sequestration. Moreover, we showed that the late-Holocene shifts in fire regime were apparently independent of changes in vegetation assemblages.

Neoglacial cooling after ~3000 cal BP may have been a principal factor in the reconstructed environmental changes. This climatic change likely negatively affected C accumulation rates. In addition, replicate centennial- to decadal-scale low peatland water table events during the Neoglacial have likely been a principal cause for the variations in vegetation assemblages, especially encountered in LLC bog, showing increasing presence of ligneous species.

To obtain a solid history of long-term C sequestration, reconstructions should take account of variations in lateral expansion of peatlands, as simple vertical rates from central records overestimate both C mass and C sequestration rates at the scale of the ecosystem. Future climate change may result in higher water tables and temperatures, which are likely to positively influence *Sphagnum* growth and thus C sequestration in this region's peatlands, while lower water tables and high temperatures may result in more important decomposition as well as peatland burning, and thus a reduced C sink function.

Keywords: peatland, carbon, Holocene, fire, paleoecology

INTRODUCTION

In the light of present and future changes in climate (IPCC, 2007), there is a general interest to improve the knowledge on links between climate and the diverse global ecosystems. Whereas past climates have been largely responsible for the present-day global distribution of vegetation zones, individual ecosystems provide feedback on climate through biogeochemical cycling (e.g. the emission of greenhouse gases). Peatland development, for instance, has resulted in a global cooling effect since the onset of the Holocene around 11 500 years before present (calendar years BP) as these ecosystems have actively sequestered carbon (C) through the uptake of carbon dioxide (CO_2) from the atmosphere (Frolking and Roulet, 2007). In order to better estimate future climate regimes, the understanding of the direction and intensity of ecosystem feedback on climate change is essential. One way to improve the knowledge on links between climate and ecosystems is through paleoclimatic and paleoecological studies. Assuming that processes and mechanisms are uniform in time, past ecosystem sensitivity and rates of change may be projected through future scenarios. Past variations are generally reconstructed using tracers or 'proxies', supposed solid indicators of past conditions.

In this thesis, variations in peatland development and C sequestration in a northern region of eastern Canada during the Holocene were studied. Considering C dynamics, peatlands are highly important ecosystems, defined by an accumulation of dead plant material of at least 30-50 cm in thickness (Charman, 2002) but possibly attaining 15 m (Clymo, 1983). Thick peat deposits, covering large surfaces in the northern hemisphere, thus contain enormous amounts of organic C, one of the constituents of CO_2 and methane (CH_4). Besides important sinks of C, peatlands are highly sensitive to variations in environmental conditions. Shifts in the delicate balance of primary production and decomposition may thus trigger important greenhouse gas disequilibrium. As past changes in climate are reasonably well quantified for the mid- and late-Holocene, studying ecosystem variations during this period may be a valuable key to future climate and ecosystem behaviour.

In northern regions, peatland development generally started after the Last Glacial Maximum with global formation culminating around 11 000-9000 years before present (BP),

resulting in a total northern peatland cover that presently attains ~4.0 million km^2 (MacDonald *et al.*, 2006 ; Yu *et al.*, 2010). These northern peatlands are generally concentrated in cool-humid temperate, and cold-humid/cold-dry boreal and (sub)arctic regions, persisting in regions with mean annual precipitation between 100 mm and 3000 mm and a mean annual temperature between -18°C and 8°C (Yu, Beilman and Jones, 2009). The global northern peatland C stock presently attains ~547 Pg (Yu *et al.*, 2010), representing about one-third of the global soil C pool. Holocene mean C accumulation rates of northern peatlands vary globally between 8.4 g m^{-2} yr^{-1} and 38.0 g m^{-2} yr^{-1} (Yu, Beilman and Jones, 2009). In (north)western Canada, where extensive boreal peatlands are present, mean C mass varies between 118-131 kg m^{-2} (Beilman *et al.*, 2008 ; Vitt *et al.*, 2000), while Holocene net C accumulation rates have been reported of 13.6-34.9 g m^{-2} yr^{-1} (Kuhry and Vitt, 1996 ; Yu *et al.*, 2003). In eastern Canada, peatland C accumulation data is much more rare, with averages roughly between 19-25 g m^{-2} yr^{-1} from southern boreal and southern, more temperate peatland (Gorham, Janssens and Glaser, 2003; Turunen *et al.*, 2004).

1. Peat accumulation dynamics

Peat accumulation generally initiates in topographic depressions with poor drainage. As peat accumulates vertically, the peatland ecosystem may expand consequently in a lateral way by paludification. A peat deposit is composed of a surface layer of living vegetation resting on litter that gradually advances to a higher state of decomposition downward and becomes submerged by the water table. Living vegetation assemblages can be spatially highly variable as a result of important variation in hydrological conditions. Nutrient availability and water table level may be the main factors determining plant composition distribution at the ecosystem scale. *Sphagnum* moss is recognized as a key factor in peat accumulation, as it actively acidifies its environment, while providing cold and nutrient-poor conditions that highly limit establishment of other plants (van Breemen, 1995). As peat accumulates, litter from the surface becomes part of the upper, (temporally) aerated layer, or acrotelm, which is eventually submerged and integrated into the anoxic catotelm.

Decomposition rates are generally high in the acrotelm, attaining a maximum near the acrotelm-catotelm boundary and get highly restricted in the catotelm (Belyea, 1996).

The large northern peatland C stock has been accumulated as primary productivity of the living vegetation generally exceeded the decomposition of underlying dead plant material over millennial timescales. Low decay rates are maintained by cold, anoxic conditions after submergence, a high acidity and a generally decay-resistant vegetation. Peat accumulation rates are thus directly linked to the water balance, or, more specifically, to the height of the water table relative to the surface, with shorter residence times in the acrotelm positively influencing net peat accumulation (Yu *et al.*, 2003). Although generally decay-resistant, plant types differ in decomposibility, with species found on dry surfaces (i.e. hummocks) being generally more resistant than wet-loving species found in hollows (Belyea, 1996). Primary production varies with roughly the same factors that determine decomposition processes, although different species react differently to changes in growth conditions (Thormann and Bayley, 1997).

As primary production and decomposition rates are highly variable, the resulting net rates of C sequestration fluctuate at diverse time- and spatial scales. Seasonal to annual variability may be linked to variations in amounts and timing of precipitation and evapotranspiration during the growing-season (Waddington and Roulet, 1999), whereas on decadal to millennial scales, broader climatic aspects and water-table driven vegetation shifts may be more important (Belyea and Malmer, 2004). Factors influencing peatland dynamics may be internally-driven (autogenic) or externally-driven (allogenic). Autogenic influence on peat accumulation may be primarily represented by relative changes in surface height and (micro)topography. Resulting from long-term changes, surface water chemistry varies, as the input of nutrients becomes limited as soon as peatland surface elevation surpasses its surroundings. During this process, peatland water tables become isolated from regional groundwater and therefore dependent on meteoric water. This shift implies changes from minerotrophic to ombrotrophic conditions and is generally referred to as the fen-bog transition (e.g. Hughes, 2000). Although this process is essentially autogenous, allogenic change may determine the timing of the transition (Hughes and Dumayne-Peaty, 2002). Furthermore, changing surface topography as peat accumulates influences drainage directions and flow rates, thereby influencing the ecosystem sensitivity to variations in precipitation.

Peatland microtopographical features as hummocks and hollows may develop as an autogenic process. Although part of a complex process, patterning is likely generated by variations in vegetation and nutrient factors and important feedback mechanisms (Eppinga *et al.*, 2008). Allogenic forcing may be subdivided in geological/geomorphological, climatic and disturbance factors. Peatland expansion is primarily related to local basin morphology as the mineral composition and the slope of the basin determine the potential for lateral accumulation of peat (Korhola, 1994 ; Mäkilä, 1997). In many regions, climate may be a principal agent as peat accumulating ecosystems are globally concentrated in regions with cool and/or humid conditions. Long-term changes in temperature and/or precipitation alter the peatland moisture balance (Barber *et al.*, 1994), yet the sensitivity of the ecosystem to climatic change may vary spatially and temporally. In other regions, sea-level change or isostatic uplift may influence peatland development, often through erosion or changes in base-level that define the peatland water table mound (e.g. Bhiry, Garneau and Filion, 2000 ; Glaser *et al.*, 2004). Finally, disturbance factors as fire may be highly important in peatland dynamics. Peatland fire is an important, natural factor in forested peatlands of the drier regions within the boreal biome, which may burn as often as upland forest fires under dry surface conditions (Turetsky *et al.*, 2004). However, open peatlands, lacking an important tree cover, may act as effective firebreaks (Hellberg, Niklasson and Granström, 2004).

2. Carbon accumulation patterns

A mass of consolidated ombrotrophic peat is typically composed of 85-95% water and 5-15% organic matter, with possibly a slight mineral fraction. Typically, organic C constitutes around 50% of organic matter mass (Beilman *et al.*, 2009 ; Borren, Bleuten and Lapshina, 2004 ; Gorham, 1991 ; Turunen *et al.*, 2001), composed of fossil plant debris, with *Sphagnum* peat generally showing slightly lower C contents than ligneous and herbaceous peat (Beilman *et al.*, 2009 ; Borren, Bleuten and Lapshina, 2004). However, extensive C content quantification over a large region indicated that C contents are largely independent of peat deposit age (Beilman *et al.*, 2009). Once the amount of sequestered C is determined and a chronology is established, rates of apparent net C accumulation can be calculated by

dividing the total C mass accumulated by the duration of accumulation (Clymo, Turunen and Tolonen, 1998). As these rates do not take account of long-term peat decomposition, deeper (and thus older) peat accumulation rates are normally underestimated relative to those of shallower deposits, even though decay in the catotelm is usually highly restricted (Clymo, 1984). Nevertheless, decreasing accumulation rates with decreasing peat age are commonly found, perhaps more important in continental climatic conditions (e.g. Beaulieu-Audy *et al.*, 2009 ; Borren, Bleuten and Lapshina, 2004 ; Vardy *et al.*, 2000 ; Vitt *et al.*, 2000 ; Yu, 2006 ; Yu *et al.*, 2003). Hence, any important allogenic influence may well overrule the effect of autogenic continuous catotelm decay. Globally, peatland C sequestration is linearly related to incident photosynthetically active radiation (PAR), showing a slightly weaker relationship with growing degree days above zero over millennial timescales (Charman *et al.*, submitted). PAR, determined primarily by latitude and cloudiness, directly influences primary productivity, and thus long-term peatland C sequestration may be driven to a greater extent by climate-driven variations in primary production than decomposition. Besides climate, peatland fires may cause an instant release of C by combustion that may attain 2.5-3.2 kg m^{-2} per fire event (Pitkänen, Turunen and Tolonen, 1999 ; Turetsky *et al.*, 2002). In addition, a postfire net C loss to the atmosphere may last for at least a decade until a local recovery of shrubs and bryophytes (Wieder *et al.*, 2009). Nevertheless, wildfire frequency in boreal peatlands is highly variable, e.g. from sites with no registered local fire since the mid-Holocene (e.g. Slave Lake; Kuhry, 1994) to a fire every 150 years (Zoltai *et al.*, 1998). Differences may be explained by regional climate, upland forest landscape connectivity and peatland hydrology and vegetation that each influence fire dynamics on different spatial and temporal scales.

3. Peatlands as archives of environmental change

Besides important stocks of organic C, peatlands potentially constitute sensitive high-resolution archives of environmental conditions (Aaby, 1976 ; Barber *et al.*, 1994), as changes in vegetation are closely linked to changes in the water balance. Under ombrotrophic conditions, the water balance is directly linked to the balance between precipitation and

evapotranspiration. More precisely, water tables are driven by variations in the summer water deficit which is most strongly correlated with precipitation (Booth, 2010 ; Charman *et al.*, 2009), although the relative importance of precipitation and temperature may vary with peatland type and setting or average moisture conditions (Booth, 2010 ; Charman, 2007). Thus, ombrotrophic peatlands incorporate a historical record of climatic variability. Peat stratigraphies generally allow for the use of multiple proxies that have a link to past water table depth (WTD) (Chambers and Charman, 2004). As vegetation types show typical optima in WTD, changes in WTD are reflected by variations in plant type abundance (Dupont, 1986 ; Bubier, Moore and Juggins, 1995), depending on the tolerance of that plant type. Peat botanical composition reconstructions (fragments >100-150 µm) or plant macrofossil analysis, provide an image of past local vegetation. This technique allows for qualitative reconstructions of WTD, however, quantitative data can be obtained by analyses of testate amoeba assemblages (Warner and Charman, 1994 ; Woodland, Charman and Sims, 1998). The tests of these micro-organisms, which exclusively live in oxic conditions near the surface, are generally well-preserved and highly abundant in *Sphagnum* peat. As vegetation types, testate amoebae assemblages are primarily determined by the depth to the water table yet these react more rapidly to changes in WTD than vegetation. This rapid response provides the opportunity to link each taxon to an accurate optimal WTD using modern training sets. As a result, subcontinental-scale transfer functions have been established that obtain a WTD for any (paleo-)testate amoeba assemblage (e.g. Booth, 2008 ; Charman, Blundell and ACCROTELM members, 2007).

The analysis of macroscopic charcoal has been used to reconstruct local fire in peatlands. Nevertheless, charcoal fragment dispersal by wind during fire and redeposition after fire may be important, and thus charcoal stratigraphies should be interpreted with caution as they may include fires from adjacent upland forest stands (Lynch, Clark and Stocks, 2004). Finally, peatlands are excellent environmental archives due to the abundance of well-preserved organic matter that facilitates radiocarbon (^{14}C) accelerator mass spectrometry dating (van Geel and Mook, 1989).

Paleoecological research results in reconstructions of environmental change for one or more records. At the core level, multiproxy research has the advantage that reconstructions from different biological, physical and geochemical proxies can be compared (Chambers and

Charman, 2004). Once an environmental history determined for a single core, replicate cores from the same or other peatlands are essential to identifty driving factors. Corresponding reconstructions from non-interacting cores indicate a large-scale, external driving factor, e.g. climate change, whereas important variability in reconstructions implies an important role of autogenic or random, local forcing. For instance, significant differences in ecosystem characteristics (e.g. peat depth, botanical composition) of two peatlands that are exposed to similar climatic conditions indicate a highly important role for local factors (e.g. basin topography or exposure to fire disturbance).

4. Peatlands in Quebec

Peatlands cover 9-12% of the Quebec territory (Payette and Rochefort, 2001). The province of Quebec ranges roughly from 45-62°N latitude, with a mean annual temperature varying between 5°C and -8°C and mean annual precipitation of 300 mm to 1300 mm, with both important north-south and east-west gradients. Due to this wide climatic range, peatland types are highly variable. The southern part of Quebec (45-54°N) is characterized by semi- and unforested raised ombrotrophic peatlands yet minerotrophic peatlands can be found as well. The northern part of the James Bay region near 54°N constitutes the biogeographical limit of ombrotrophic peatlands, as *Sphagnum* growth becomes limited due to cold conditions. The subarctic region, from the southern limit of the Hudson Bay northwards, is characterized by palsa (plateau) peatlands (Payette and Rochefort, 2001).

Although millennial- to decadal-scale reconstructions of deforestation and fire frequency, with some linkages to climate, are numerous in boreal Quebec (e.g. Bergeron and Archambault, 1993 ; Carcaillet *et al.*, 2001 ; Filion, 1984 ; Payette, Gauthier and Grenier, 1986), no research on peatland C sequestration patterns had been performed before the DECLIQUE research program started at Université du Québec à Montréal. Holocene peatland development has been studied in the forest-tundra (Bhiry, Payette and Robert, 2007) and the lichen woodland biomes (Beaulieu-Audy *et al.*, 2009), whereas peatland development and C accumulation rates are available from the Eastmain region in western Quebec for the late-Holocene (Loisel and Garneau, 2010).

The Eastmain region is characterized by frequent and large-surface burning with a mean frequency of a fire each 90-100 years (Mansuy *et al.*, 2010 ; Payette *et al.*, 1989). This optimum in eastern Canadian forest fire regimes is probably the result of a combination of relatively low-humidity climatic conditions and the dominance of a closed-crown coniferous forest that favours fire continuity. Due to an important occurrence of forest fires, this region's peatland C sequestration may have been highly influenced by local burning during the Holocene.

Eastern Canada Holocene climate history has been characterized by a mid-Holocene thermal maximum (~4000 cal BP; Kaufman *et al.*, 2004 ; Viau *et al.*, 2006). This warm period was delayed relative to other regions of boreal North America due to the proximity of the residual Laurentide Ice Sheet. A consecutive cooler period, referred to as the Neoglacial, possibly lasted for several millennia that caused large-scale forest cover regression and induced permafrost occurrence in subarctic peatlands (Bhiry, Payette and Robert, 2007 ; Kerwin *et al.*, 2004 ; Payette and Gagnon, 1985).

5. Aims

The Eastmain region occupies a particular position in the northern boreal climate domain, being located toward the cold-wet extreme that is poorly represented by the currently available data on C sequestration (Yu, Beilman and Jones, 2009). In addition, knowledge on peatland fire dynamics is largely based on reconstructions from central and western boreal Canada. In eastern Canada, however, differential past climate regimes may have caused different historical burning patterns. In order to obtain a better image of the link between peatland fire and C dynamics and explore the gradient of peatland types susceptible to burning, reconstructions from open peatlands as found in the Eastmain region are essential. Hence, this thesis focuses on the relationship between C sequestration and fire regimes. As fire and C sequestration are expected to be closely linked to vegetation dynamics and water table heights within the regional climatic context, these aspects were included in the project as well. This study encompasses data from multiple cores sampled in three peatlands to cover

possible spatial variability in the reconstructions and to be able to identify driving factors and their respective spatial scale.

Specifically, the aims of this thesis include:

1) A quantification of total amounts of C sequestered in ombrotrophic peatlands in the Eastmain region of northern Quebec, Canada. As quantifications of peat C stock at the ecosystem scale and mean C mass at the scale of the region are globally rare, and even absent for eastern Canada, these data will provide more accurate estimations of northern Quebec C stocks;
2) A reconstruction of variations in Holocene C accumulation rates in these peatlands, including both temporal and spatial variations in C sequestration for peatlands individually. An accurate quantification of variations in vertical accumulation and lateral expansion will show the respective importance of autogenic and allogenic factors;
3) The establishment of the links between reconstructed C accumulation rates and past variations in local vegetation and water table depth. As water table depths may be linked to climate variations, Holocene hydroclimatic variations are quantified;
4) The establishment of the link between reconstructed C accumulation rates and past peatland fire regimes. Moreover, driving factors and scales for past fire patterns are identified.

At large, the mechanisms derived from the results of this thesis will contribute to more accurate scenarios of future environmental change (e.g. climate and fire) on C sequestration in regions with comparable environmental settings.

Quantification of total C sequestered in Eastmain region peatlands and reconstructions of past accumulation rates will be discussed in Chapter 1. For each of three peatlands, areal extent, surface topography, basin morphology and peat thickness were measured and linked to mean C density values to obtain an ecosystem C stock. Chronologies from multiple cores from each peatland were merged to create an ecosystem-specific age-

depth model. These models were used to infer patterns of vertical peat accumulation and lateral expansion.

The influence of core-scale ecohydrological change on C accumulation rates is the principal subject of Chapter 2. Reconstructions of C accumulation rates and ecohydrology were obtained from cores sampled in the deepest section of each peatland. Holocene rates of C accumulation were quantified by age-depth modelling and peat C density analyses. Ecohydrological changes were reconstructed by plant macrofossil and testate amoeba analyses. A transfer function was applied to the past amoeba assemblages to infer Holocene fluctuations in local water table.

Chapter 3 focuses on possible linkages between historical peatland fire and C sequestration at the core scale. Both C accumulation rates and fire records were quantified from lateral cores. Past fire events were identified from stratigraphic, macroscopic charcoal and a threshold was used to separate local from regional fires. Long-term variations in both fire intensity or severity and fire frequency were related to past C accumulation rates.

Finally, results are briefly summarized in the conclusion, and the importance of factors considering C accumulation dynamics are discussed with respect to the spatial and temporal scales considered.

CHAPTER I

QUANTIFYING SPATIAL AND TEMPORAL HOLOCENE CARBON ACCUMULATION IN OMBROTROPHIC PEATLANDS OF THE EASTMAIN REGION, QUEBEC, CANADA

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Article published in Global Biogeochemical Cycles.

Résumé

Les tourbières nordiques constituent des réservoirs importants de carbone (C) organique. La dynamique du C dans ces tourbières influence de façon importante la concentration atmosphérique des gaz à effet de serre et devrait donc être considérée dans l'étude des changements climatiques futurs. La quantification des taux d'accumulation du C dans les tourbières est souvent basée sur des résultats d'analyse d'une carotte centrale. Cependant, les résultats de ces carottes centrales peuvent surestimer les taux d'accumulation s'ils sont extrapolés à l'échelle de l'écosystème. Nous proposons ici une reconstruction de l'accumulation du C à l'échelle de la tourbière pour trois tourbières ombrotrophes du Québec boréal à l'aide de plusieurs carottes prélevées dans chacune d'elles.

Dans cette étude, les stocks de C ainsi que les variations temporelles en taux d'accumulation ont été quantifiés. De multiples datations au ^{14}C sur plusieurs carottes par tourbière ont servi à la création de modèles d'accumulation représentatifs de chaque écosystème.

Le début de l'accumulation de tourbe date d'environ 7500 cal BP, soit peu de temps après la déglaciation de la région. L'accumulation verticale de tourbe a diminué au cours de l'Holocène, tandis que l'expansion a été rapide durant les premiers millénaires pour ensuite ralentir vers l'Holocène moyen. L'accumulation du C à l'échelle de chaque écosystème a culminé entre 5250 et 3500 cal BP, tandis que le taux moyen holocène est de 16,2 g m^{-2} an^{-1}.

Les tourbières de la région d'Eastmain semblent constituer des puits de C plus modestes que celles de l'Alaska, de l'ouest canadien ou de l'ouest de la Sibérie. Cependant, des différences méthodologiques empêchent une juste comparaison de ces accumulations. La période caractérisée par la plus grande quantité de C rajoutée à l'écosystème correspond à une période de ralentissement à la fois de l'expansion latérale et de l'accumulation verticale. Ce ralentissement de l'accumulation du C à l'Holocène récent a été attribué à des facteurs autogènes et allogènes. Les diminutions de l'accumulation de C dans les tourbières étudiées pourraient avoir été causées par un abaissement autogène des nappes phréatiques ou encore par le refroidissement du Néoglaciaire. Les résultats de cette étude montrent que les reconstructions de l'expansion de la tourbière à l'Holocène sont essentielles afin d'obtenir des quantifications précises de l'accumulation à l'échelle des tourbières.

Abstract

Northern peatlands represent important stocks of organic carbon (C). Peatland C dynamics have the potential to influence atmospheric greenhouse gas concentrations and are therefore of interest concerning future climate change. Quantification of Holocene variations in peat C accumulation rates is often based on a single, deep core. However, deep cores may overestimate accumulation rates when extrapolated to the ecosystem scale. We propose a reconstruction of C sequestration patterns based on multiple cores from three ombrotrophic peatlands in boreal Quebec, Canada.

Both total C accumulation and temporal variations herein were quantified. Radiocarbon-dated stratigraphies from different sections resulted in peatland-specific age-depth models.

Peatland initiation started rapidly after deglaciation around 7500 cal BP. Vertical accumulation slowed down in the course of the Holocene, whereas lateral expansion was rapid in the early stages but slowed down near mid-Holocene. Total C accumulation showed maximum rates between 5250 and 3500 cal BP with a regional mean Holocene apparent rate of 16.2 g m^{-2} yr^{-1}.

The Eastmain peatlands have been modest sinks of organic C compared to those of Alaska, western Canada and western Siberia, although differences in calculation methods hamper direct comparisons. Considering within-peatland dynamics, maximum total C sequestration coincided with a period of slowing down in both lateral expansion and vertical accumulation. Late-Holocene diminishing peatland C sink functions have been attributed to autogenic as well as allogenic factors. Height-induced surface drying and/or Neoglacial cooling effects may have forced the slowing down of C sequestration in the studied bogs. Results further show that, in order to obtain an accurate quantification of past C sequestration, reconstructions of peatland expansion are essential.

1.1. Introduction

At a local scale, boreal peatlands generally act as small sinks of carbon dioxide (CO_2) and large sources of methane (CH_4), while on a global scale they constitute a net sink of organic carbon (C) (Roulet, 2000). Greenhouse gas fluxes between peatlands and the atmosphere have the potential to influence climate radiative forcing on millennial timescales (Frolking and Roulet, 2007) and thus peatland C dynamics are of interest considering their feedback on atmospheric global warming. Peatland influence on atmospheric composition is illustrated by the fact that global Holocene trends in peatland initiation and expansion are linked with atmospheric CO_2 and CH_4 concentration (Korhola *et al.*, 2010 ; MacDonald *et al.*, 2006).

Carbon accumulates as organic detritus under waterlogged conditions when the rate of biomass production exceeds its decomposition (Turunen *et al.*, 2002), with both processes showing variations on seasonal to millennial timescales (Beilman *et al.*, 2009 ; Borren, Bleuten and Lapshina, 2004 ; Bubier *et al.*, 2003 ; Dorrepaal *et al.*, 2009 ; Schimel *et al.*, 2001). The global northern peatland store has been accumulated notably for 12 000-8000 years (MacDonald *et al.*, 2006), resulting in a C stock of ~547 Pg covering an area of 3-4 × 10^6 km^2 (Yu *et al.*, 2010).

Global peat accumulation patterns vary with latitude (Beilman *et al.*, 2009), permafrost presence (Turetsky *et al.*, 2007 ; Beilman *et al.*, 2009), precipitation (Gorham, Janssens and Glaser, 2003), continentality (Tolonen and Turunen, 1996), peatland age and depth (Belyea and Clymo, 2001 ; Clymo, 1984), fire events (Pitkänen, Turunen and Tolonen, 1999 ; Turetsky *et al.*, 2002), peatland type (Turunen *et al.*, 2002) and surface microtopography presence (Eppinga *et al.*, 2009 ; Malmer and Wallén, 1999 ; Swanson, 2007). Hence, the characterization of regions with important peatland cover is essential to accurately quantify the global peat C store and estimate future greenhouse gas budgets.

Considering accumulation gradients at a local scale, autogenic changes in hydrology and differential peat growth can be linked to changes in peatland surface topography (Belyea and Baird, 2006), while spatial variations in the C balance on short and long timescales may be considerable (Malmer and Wallén, 1999 ; Waddington and Roulet, 1999). In topographic depressions with a uniform shape, peat accumulation is typically initiated in the lowest

section which implies that subsequent peat accumulates both vertically and laterally. Hence, the age of peat inception generally decreases toward the basin margins and thicker peat deposits are located near the center of the basin. In most peatland types, margins differ hydrologically and botanically from the central parts. Margins are often characterized by a denser tree and shrub cover as water tables are lower and more sensitive to fluctuations (Bauer *et al.*, 2009 ; Bubier, 1991).

Current variable estimations of the northern peatland C stock are the result of differing assumptions on global average peat depth (Gorham, 1991) and bulk density (Turunen *et al.*, 2002) or inaccurate spatial inventories and volume models (Beilman *et al.*, 2008 ; Vitt *et al.*, 2000). Apparent rates of C accumulation (expressed in g m^{-2} yr^{-1}) calculated from a single, central core within a peatland may result in estimations of total C sequestration that inadequately represent the entire system (Turunen *et al.*, 2002). One reason for central coring sites might be a preference for long historical records of peat development. However, this record might poorly represent the entire ecosystem as spatial variability can not be quantified. Poor temporal correlation between vertical and lateral peat accumulation within a single bog has been found repeatedly (e.g. Korhola, 1994 ; Korhola *et al.*, 2010). Hence, more accurate C accumulation rates can be obtained using patterns from various downcore sections taking account of variations in depth and minimal ages of peat inception, especially if local basin topography is complex.

Spatial peat accumulation reconstructions and quantifications of C sequestration have been performed in Scandinavia (Korhola, 1994 ; Korhola *et al.*, 1996 ; Mäkilä, 1997 ; Mäkilä and Moisanen, 2007), Scotland (Chapman *et al.*, 2009), western Canada (Bauer, Gignac and Vitt, 2003 ; Beilman *et al.*, 2008) and western Siberia (Borren and Bleuten, 2006 ; Sheng *et al.*, 2004 ; Turunen *et al.*, 2001). Despite an important coverage on the Quebec territory (9-12%, Payette and Rochefort, 2001), C sequestration data from boreal peatlands east of Hudson Bay and James Bay is limited. Presently, C accumulation rates have been quantified by Loisel and Garneau (2010) from four one-meter long cores collected in two of the Eastmain peatlands presented here. Peat and C accumulation rates have also been determined from cool-continental and maritime bogs in the southern regions of eastern Canada (Turunen *et al.*, 2004) as well as from ombrotrophic peatlands in the mixedwood forest biome (Gorham, Janssens and Glaser, 2003 ; Turunen *et al.*, 2004). Beaulieu-Audy *et al.* (2009)

calculated vertical peat accumulation rates in boreal peatlands from the La Grande Rivière region, northern Quebec, but did not provide quantitative data on the rate of C accumulation.

In this paper we present reconstructions of peat accumulation and long-term C sequestration patterns from three pristine ombrotrophic peatlands in the Eastmain region of northeastern Canada (51°50'-52°20'N/75°00'-76°00'W). The main objective is to quantify regional Holocene C accumulation in terms of C mass, density and accumulation rates. A secondary objective consists in increasing the knowledge on C sequestration variability within and between peatlands from the same region and to compare our results with data from western North America and Eurasia.

Previous research on lateral and vertical peat accumulation has shown rapid lateral expansion of peatland ecosystems in the early stages, with a slowdown in the course of their development (Mäkilä, 1997 ; Mäkilä and Moisanen, 2007), although smaller-scale variations in peatland expansion rates may be primarily controlled by basin topography (Korhola, 1994). In accordance, we expect slowing down of peatland expansion in the Eastmain region toward the late-Holocene. However, global tendencies in vertical peat accumulation may be less uniform. Peatlands in oceanic settings typically show concave age-depth models, resulting principally from the effects of constant productivity and continuous C loss in the catotelm as indicated by Clymo (1984). In contrast, continental bogs often show convex age-depth models (Kuhry and Vitt, 1996 ; Turunen *et al.*, 2001), implying long-term slowdown of accumulation during peatland development (Yu *et al.*, 2003). This trend may be driven by autogenic or allogenic influence or a combination of both (Belyea and Malmer, 2004). As the Eastmain region has neither an oceanic nor a strictly continental setting, we hypothesize that the studied peatlands show approximate linear vertical accumulation.

Holocene C accumulation rates of two boreal peatlands located ~400 km south of our study region were quantified at approximately 21.9 and 29.4 g m^{-2} yr^{-1} (Gorham, Janssens and Glaser, 2003). As global C sequestration rate optima are associated with a mean annual temperature (MAT) of around 0°C (Beilman *et al.*, 2009) or 0°C to 2.5°C (Yu, Beilman and Jones, 2009), potential rates are likely to decrease northward of ~50°N in Quebec. The Salym-Yugan peatland complex in western Siberia is subjected to MAT and mean annual precipitation (MAP) comparable to the Eastmain region; showing a mean rate of 17.2 g m^{-}

^{2} yr^{-1} (Turunen *et al.*, 2001). Based on these data, we hypothesize that Holocene C accumulation rates of the Eastmain region peatlands might average 15–25 g m^{-2} yr^{-1}.

1.2. Study region

The three studied peatlands Lac Le Caron (LLC), Mosaik (MOS) and Sterne (STE) are located in the Eastmain river watershed in the boreal forest region of the James Bay lowlands of Quebec (Fig. 1.1). Regional MAT is -2.1 ± 0.2°C (January: -22.0 ± 0.5°C; July: 14.6 ± 0.2°C) and MAP is 735 ± 12 mm, of which about one third falls as snow (Hutchinson *et al.*, 2009). The region is characterized by Proterozoic bedrock and glacial and postglacial landforms as drumlins and eskers. Deglaciation occurred between 8500 and 7900 cal BP (Dyke, Moore and Robertson, 2003) and was followed by the Tyrrell Sea invasion that caused the deposition of marine and deltaic sediments in the western part of the territory. The regional upland vegetation corresponds to the limit of the *Picea mariana*-feathermoss and *Picea mariana*-lichen bioclimatic region (Saucier *et al.*, 1998). Peatlands cover ~7% of the Eastmain region varying from treed bogs to wet fens (Grenier *et al.*, 2008). The studied peatlands are classified as pristine (eccentric) raised bogs characterized by a well-developed hummock-hollow patterned surface with deep pools (~2 m) in their central sections (Loisel and Garneau, 2010). Peat accumulation started by paludification as peat types identified at the base of the cores did not reveal past infilling ponds.

Lac Le Caron bog (LLC) (52°17'15"N/75°50'21"W; 2.24 km^2 area) is located in the northwestern part of the region (Fig. 1.1). The western part of the basin is bordered by a steep, ~40 m-high escarpment while the eastern limit is relatively flat with a stream flowing southward. The center of LLC bog is treeless with wet hollows and large pools. Sedges are more abundant than in the surrounding ribbed section. The bog margins are forested and dominated by *Picea mariana*. A fen characterized by a near-surface water depth is present next to the pool sections and sparse *Larix laricina* grow occasionally on lawns. Drainage is directed toward the eastern section of the system. The mineral basin sediments are highly variable from fine sand to silt.

Mosaik bog (MOS) (51°58'55"N/75°24'06"W; 2.67 km^2 area) is located 45 km southeast of LLC bog (Fig. 1.1), where topography is less pronounced. The center of this peatland is characterized by an important presence of wet hollows and numerous large pools with outcrops present in the northern section. Bog margins are colonized by *Picea mariana* while the southwestern part was affected by a local fire in 1997 and is characterized by sparse *Pinus banksiana* Lamb. Multidirectional hummock-hollow patterns may indicate a complex pattern of drainage. Mineral sediments range from coarse to fine sand with pebbles or (bed)rock.

Finally, Sterne bog (STE) (52°02'37"N/75°10'23"W; 1.72 km^2) is located 17 km northeast of MOS bog (Fig. 1.1). As MOS bog, its central section is very wet with many large pools. The eastern part has an indistinct forest-bog transition. In the southwestern part, a small stream separates the bog from a large poor fen. The mineral basin of STE bog is characterized by poorly sorted coarse and fine sands and a frequent presence of pebbles or (bed)rock.

Each of the studied peatlands shows a forested border of varying width. This ecotonal limit shows both abrupt and indistinct transitions. On the open peatlands, trees of *Picea mariana* are sparsely distributed on hummocks. Ericaceous shrubs such as *Chamaedaphne calyculata*, *Kalmia angustifolia*, *Rhododendron groenlandicum* and *Andromeda glaucophylla* are distributed following a moisture gradient. Cyperaceae are abundant in lawns and hollows: *Eriophorum vaginatum*, *Trichophorum cespitosum* and *Carex* spp. Dominant bryophytes are *Sphagnum fuscum* and *Sphagnum angustifolium* on hummocks, *Sphagnum russowii* and *Sphagnum magellanicum* on lawns, whereas *Sphagnum cuspidatum*, *Sphagnum fallax* and *Sphagnum majus* are frequent in wet hollows.

1.3. Material and methods

1.3.1. General approach

Quantifying C accumulation in a peatland implies the integration of areal extent, variability of deposit thickness and related C density. A chronological approach integrating

both age and peat thickness was adopted to reconstruct rates of C sequestration through time. The accuracy of the reconstruction depends directly on the spatial uniformity of the age-depth relation. Models of peat cover thickness were created from field probing measurements and Ground-Penetrating Radar (GPR) analyses. Added to these models were radiocarbon datings from multiple cores sampled in different sections of the peatland resulting in age-depth relationships. Afterward, peatland development was divided into 250-year time slices that were linked to peat depth values. The combination of peat depths and the spatial cover thickness model resulted in a definite volume of peat accumulated during each time slice. This volume was then converted in mass of organic C using bulk density and loss-on-ignition (LOI) data, resulting in a quantification of the C flux. The total mass of C was represented by the sum of C fluxes for all time slices, while lateral expansion rates (expressed as $m^2\ yr^{-1}$) were calculated from the increase in surface area divided by the duration of the period considered.

1.3.2. Peat depth models

The areal extent of the studied peatlands was determined by aerial photo interpretation and field validation. Peatland surface altitude was obtained by a Trimble 5800/5700 Differential Global Positioning System (DGPS) along a number of transects per peatland during the summer of 2006. Peat thickness was determined by a combination of manual probing with an Oakfield soil sampler and GPR analyses. Survey points were located along grids (probing) or transects (GPR) localized using DGPS (Fig. 1.1). At each sampling point, the composition of underlying mineral material was described. Manual probing was realized at 100- and 200-m intervals. GPR measurements were performed with a PulseEKKO (Sensors & Software Inc.) at 0.25 m to 1 m resolution using both 100 MHz and 200 MHz antennae during a winter campaign in 2007 and the summer of 2008. Time delay between electromagnetic wave emission and reception was converted to peat cover thickness using a mean peat velocity, determined by common mid-point analysis (Neal, 2004) and target-to-depth technique (Rosa *et al.*, 2009). Data were processed with basic editing (Dewow filter, AGC gain, FK migration) (Jol and Bristow, 2003) with Reflexw software (Sandmeier, 2005)

in order to identify the organic-mineral contact. Data were compiled to create peat thickness models using ArcGIS 9.3. Peat thickness models were created using ordinary kriging interpolation using a spherical model to fit to the variograms. Model selection was based on the lowest mean standardized error obtained by cross-validation, as in most cases mean root-mean-square errors were not significantly different (F test, $P = 0.05$). Cross sections showing present-day peatland surface and mineral basin topography were created to obtain an image of peatland geometry. As the cross-section was supposed to deliver a global image, surface topography data was smoothed by a locally weighted least squared error method with 5% smoothing factor using Kaleidagraph 3.6.

1.3.3. Age-depth relationships and peat volume accumulated per period

Multiple cores were extracted from each peatland during field campaigns in August 2006 and July 2007 (Fig. 1.1). Based on manual probing results, a core was sampled where peat thickness was found to be maximum. Although referred to as central core, its position did not correspond to the geographic center of the peatland because of the complexity of the basin topography. In addition, five to six shallower lateral (L) and transitional (i.e. forest-bog transition: T) cores, located along the margins, were extracted from each peatland (Fig. 1.1). L- and T-cores were sampled in different quadrants of the peatland to cover spatial variability. All profiles were collected using a Box corer (10×10 cm width) for the top 1 m and Russian peat samplers (4.5-cm or 7.5-cm diameter) for the deeper, compacted peat. Cores were extracted from surface lawn microforms as these are likely to be more sensitive to environmental change than hummocks (Nordbakken, 1996 ; Rydin, 1993). Monolith lengths ranged from 64 to 483 cm with variability within and among peatlands. Cores were wrapped in plastic and covered by PVC tubes before storage at 4°C until analysis. In the laboratory, cores were cut into 1-cm thick slices and subsampled for analysis. Five cores per peatland were investigated in order to reconstruct peat-forming vegetation assemblages (Troels-Smith, 1955). To obtain reliable chronologies, a total of 91 subsamples were radiocarbon dated at the Keck-CCAMS facility (Irvine, USA) and Beta Analytic Inc. (Miami, USA). For each peatland, chronologies of the deep core and two lateral cores were based on numerous

datings, whereas of the remaining lateral cores only basal peat was dated (Table 1.1). If present, *Sphagnum* stems were dated because these yield most reliable ^{14}C dates (Nilsson *et al.*, 2001). Other levels were dated using leaf or seed fragments of Ericaceae and Cyperaceae, and in some cases charcoal fragments. Datings were calibrated using the IntCal04 calibration curve (Reimer *et al.*, 2004) within the Bchron software package (Haslett and Parnell, 2008). All ages are expressed as calendar years before present (BP = before AD 1950).

For each peatland, Bchron output ages were converted to an age-depth model using JMP 7.0. The surface was assigned either -56 or -57 cal BP (i.e. 56 years after reference year AD 1950), as the peat cores were sampled in AD 2006 and 2007. Only the permanently waterlogged catotelm peat was modelled, as decay is important in the (temporally) aerated acrotelm resulting in differential accumulation dynamics. Both linear and polynomial models were tested, as catotelm peat accumulation rarely conforms to a simple linear relation between age and depth (Blaauw and Christen, 2005). The original model showed heteroscedasticity, hence a transformation was applied by modelling the square root of the original dependent variable "depth". Parameters were tested and the distribution of residuals was studied for each model. After selection of the appropriate age-depth model, the peat sequence history was divided into 250-year time slices. As the acrotelm was excluded, the upper limit of the model was defined at 250 cal BP. Concerning the base of the peat cover, an additional assumption had to be made. GPR measurements performed after coring detected sections of the peatland containing peat deposits of which the thickness exceeded the length of the central cores. As a result, the deepest sections of the peatland are not covered by the age-depth model. As extrapolation of age was assumed to be unreliable due to nonlinearity of age-depth relationships, the adopted chronologies only cover the range corresponding to that of the central core. Levels that exceeded this depth were assigned an age range of "*x*-8000 cal BP", with *x* representing the age of the oldest dated sample for each peatland. The upper and lower limits of each time slice were linked to a depth determined from the corresponding age-depth model. These depths were integrated into the peat thickness model to estimate the volume of accumulated peat per time slice. Volumes were calculated using the 3D Analyst toolbox in ArcGIS 9.3.

1.3.4. Peat organic C content and density

The organic C content of peat was calculated using data from LOI analysis and mean C content of organic matter (OM). The product of bulk density and LOI analyses determined the density of OM (Dean, 1974). Dry bulk density was measured from consecutive 1 cm^3 subsamples after drying in an oven for 16 hours at 105°C. Subsamples were combusted at 550°C for 3.5 hours to determine LOI (Heiri, Lotter and Lemcke, 2001) and the resulting OM density was converted to C mass per unit volume (C density) assuming a constant mean peat C content of 50% relative to OM. Subsequently, mean C density values for each peatland were applied to the undecomposed, upper peat (younger than 250 cal BP) and decomposed, deeper peat. The boundary between upper and deeper peat corresponded closely to the level of C density culmination. Considering the deep peat, C density showed high variability both within cores and peatlands. As a uniform relationship between peat depth and C density was absent within the deep peat, we applied a mean C density for each peatland individually. This mean density for deep peat was based on subsamples of central and lateral cores within each peatland.

1.3.5. Peat C stocks and C accumulation rates per period

The amount of C added per time slice was calculated by multiplying the volume of peat by the mean C density. The total C mass of each peatland was obtained by the sum of C accumulation values for all time slices. Holocene C accumulation rates were calculated by dividing the mass of C accumulated (g) by the mean period of accumulation (yr) for a constant surface area (m^2) (Clymo, Turunen and Tolonen, 1998). This mean period of accumulation was based on the peat depth distribution and the peatland-specific age-depth model. In addition, recent apparent rates of C accumulation (*sensu* Tolonen and Turunen, 1996) were calculated for the upper peat layer (i.e. the most recent time slice).

1.4. Results

1.4.1. Basin and surface topography

The stratigraphic mineral-organic transition was generally sharp in each peatland, identified through GPR images, LOI results as well as from probing. Peat thickness values from manual probing and GPR at identical locations were highly linearly correlated ($r^2 = 0.90$; n = 30) showing the accuracy of both methods. Applied mean velocity was relatively high at 0.040 and 0.046 m ns^{-1} (Hänninen, 1992 ; Leopold and Volkel, 2003 ; Rosa *et al.*, 2009) possibly due to winter conditions and low water temperatures.

The basin topography shows high small-scale variability possibly associated with local presence of bedrock or boulders (Fig. 1.2). Especially in LLC bog, local depressions are present, indicating that the peatland may have been formed by the fusion of numerous mesotopes (*sensu* Charman, 2002).

Present-day surface topography shows variation in altitude of 2 to 8 m within the peatlands (Fig. 1.2). LLC (sloping southeastward at ~5 m km^{-1}) and STE (sloping westward at ~3 m km^{-1}) have most pronounced gradients, whereas MOS bog has a flatter surface. However, MOS and STE have more typical raised bog shapes with an elevated center, while LLC bog does not show the typical dome morphology.

1.4.2. Peat cover thickness

Maximum peat thickness is 531 cm in LLC bog as obtained by probing, showing that the deepest core (LLC_C), measuring 483 cm, was effectively sampled within the thickest peat deposits. However, GPR analyses in MOS and STE bogs show maximum peat deposit thickness of 385 cm and 412 cm, whereas their central cores measured 297 cm (MOS_C) and 286 cm (STE_C). Generally, the thickest peat deposits are located off-center within each peatland, influenced by local basin topography (Fig. 1.2). Because the dated peat cores did not cover the complete range of peat thickness, the age-depth models of MOS and STE bogs

are slightly biased toward the shallower peat deposits and do not represent basal peat accumulation as well as does the LLC bog chronology.

1.4.3. Radiocarbon dating and age-depth models

Basal ages of the different lateral and central cores range from 1211 to 7520 cal BP (Table 1.1). The basal ages of the three studied peatlands are 7520 cal BP for LLC bog, 7340 cal BP for MOS bog and 7127 cal BP for STE bog and confirm that peat inception started early after deglaciation in the region (Dyke, Moore and Robertson, 2003). For LLC bog, a second-degree polynomial model ($r^2 = 0.86$) with significant parameters ($P < 0.05$) best fitted the relationship between the square-root of depth and age, based on the distribution of the residuals (Fig. 1.3). Although age-depth relationships were modelled with a transformed "depth" variable, the models are shown with linear axes. MOS ($r^2 = 0.75$) and STE ($r^2 = 0.69$) bogs were best represented by a significant linear model (both $P < 0.0001$).

The convex model representation of each peatland shows vertical peat accumulation slowdown in the course of its development. The most important change in accumulation rate occurred in LLC bog, declining from 0.103 cm yr^{-1} between 7520 and 6520 cal BP to 0.016 cm yr^{-1} between 1250 and 250 cal BP. MOS bog accumulation slowed down from 0.036 to 0.018 cm yr^{-1} whereas STE bog rates showed a decline from 0.034 to 0.019 cm yr^{-1} both from the first millennium to the 1250-250 cal BP period. Highest similarities between age-depth models are found toward the late-Holocene.

1.4.4. C density and mass per unit area

The mean C density from all peat samples is 44 kg m^{-3} (SE = 0.0003; n = 3606) with slight variations among peatlands (Table 1.2). C density was lowest in the living moss layer and generally increased downward until a peak around 250 cal BP (Fig. 1.4). C density did not show a consistent increasing trend toward deeper peat in the anaerobic section (Fig. 1.4), indicating that vegetation type-related humification might be as important as the age of peat

formation. Instead, temporal variations in microforms and peat-forming vegetation may explain the variations in the deeper peat C density data. Stratigraphic analysis (Troels-Smith, 1955) showed variations in peat-forming vegetation within each core with alternance of *Sphagnum*, herbaceous and ligneous peat, which may be due to internal microform dynamics or external forcing, or both (Fig. 1.5).

Mean area-weighted C mass per unit area is 91 kg m^{-2} with a highest mean in LLC bog (Table 1.2). As LLC does not show the highest mean C density, its higher C mass per unit area primarily results from a higher mean peat thickness (Table 1.2). Thus, despite a smaller surface area than MOS bog, LLC bog has been the most important sink of organic C throughout the Holocene.

1.4.5. Peatland area and C accumulation reconstructions

Lateral peatland expansion rates were high in the early stages of development (Fig. 1.6). Early rapid peatland development was followed by a gradual decline in lateral accumulation rates that started as early as 6000 cal BP in the three peatlands. Although initial paludification is suspected to have a maximum age of 8000 cal BP, the combination of age-depth and spatial peat thickness modelling implies that 50% of the present peatland area was covered by peat deposits around 5500 cal BP.

Ecosystem-scale C flux reconstructions show similar trends for each peatland (Fig. 1.7), with an increase during the first millennia, peaking during the mid-Holocene, followed by a decline toward the late-Holocene. The most recent time-slice (0-250 cal BP), which includes the acrotelm, shows high values as decay is incomplete. Despite the comparable tendencies between the three peatlands, variations in timing are visible (Fig. 1.7). In LLC bog, C fluxes show greater variations in time, culminating between 5250 and 5000 cal BP, with an equivalent rate of 41 400 kg yr^{-1}. In MOS and STE bogs, C accumulation appears to have been more stable, with highest fluxes of 27 000 kg yr^{-1} and 18 000 kg yr^{-1} between 3750 and 3500 cal BP, respectively. Lateral expansion has been most important in the early stages of peat bog development, while vertical accumulation of peat slowed down during the entire history of the peatlands (Fig. 1.3 and 1.6). However, the

periods with maximum ecosystem C flux (5250-5000 cal BP in LLC bog and 3750-3500 cal BP in MOS and STE bogs) correspond to the optima in the balance of both directions of accumulation. Hence, although counterintuitive, maximum ecosystem C fluxes coincided with periods of diminishing lateral and vertical accumulation rates in each of the ecosystems.

Although MOS bog presently covers the largest area, LLC bog has accumulated the highest amount of C: 242×10^6 kg, compared to 217×10^6 kg C for MOS bog. The smaller STE bog presently contains 149×10^6 kg C (Table 1.2). Taking into account peatland surface area and mean age of peat initiation, Holocene C accumulation rates were 18.9, 14.4 and 15.2 g m^{-2} yr^{-1}, for LLC, MOS and STE bogs respectively, corresponding to an area-weighted regional mean (±SE) of 16.2 g m^{-2} yr^{-1} (±1.4). Mean C accumulation rate of the most recent time slice is 56.4 g m^{-2} yr^{-1} (±1.7); values for peatlands individually are shown in Table 1.2.

Results from peat composition analysis show that ombrotrophication occurred rapidly after initial peat accumulation in a number of cores with some sections showing delayed or very recent shifts to ombrotrophic conditions, e.g. MOS_L1 (~160 cal BP) and STE_L4 (~220 cal BP) (Fig. 1.5). Nevertheless, no significant differences in mean age of ombrotrophication are discernible between ecosystems ($F = 0.91$; $P = 0.45$).

1.5. Discussion

1.5.1. Age-depth modelling and C accumulation patterns

The age-depth models presented are based on chronologies from cores that covered the variability in deposit thickness of each peatland. The age-depth model of LLC bog represents well the range of peat cover thicknesses and has a high r^2 of 0.86. However, MOS and STE bog chronologies are slightly biased toward younger peat deposits as the obtained chronology did not cover the deepest peat. In addition, their respective age-depth models show lower r^2 at 0.75 and 0.69 which implies that the relationship between age and depth is more variable within these peatlands. For these reasons, the older sections of the MOS and STE age-depth model need to be interpreted with caution. It is probable that a central core

within the deepest section of these peatlands results in a more convex age-depth model, revealing C accumulation patterns that would more closely resemble those of LLC bog.

The presented reconstructions of the three peatlands show that lateral expansion has been important in their early development. In addition, net vertical accumulation rates have diminished continuously, causing a slowdown of C accumulation rates. Although trends are comparable, MOS and STE bogs showed less variable rates of vertical accumulation than LLC bog, which resulted in a minor decrease in C accumulation rates in the late-Holocene. Generally, peatland hydrology and microhabitat patterns are influenced by both internal dynamics (autogenic factors) and external forcing (allogenic factors) on varying timescales (Belyea and Baird, 2006 ; Payette, 1988), which will be discussed here relative to the observed patterns.

1.5.2. Autogenic factors

Internal processes may explain to some extent different tendencies in peatland development. Of all types of autogenic change, ombrotrophication may be the most important considering peatland C sequestration (Charman, 2002). Typical net C accumulation rates differ between fens and bogs (Turunen *et al.*, 2002 ; Yu, 2006) and thus hydroseral succession may well result in a step-like change in C cycling (Belyea, 2009). However, stratigraphic analyses have not shown distinct differences in the timing of ombrotrophication between the studied peatlands. In contrast, Loisel and Garneau (2010) reported more recent ombrotrophication in MOS bog in comparison with LLC bog. However, only one section of each peatland was considered in their study.

The uniform highly sloping surface (~5 m km^{-1}) of LLC bog (Fig. 1.2) may have resulted in a more effective lateral drainage through subsurface flow than in MOS and STE bogs. This could have resulted in a more frequent peat surface drying since slope development. An effective drainage in LLC bog is also visible through the minor presence of wet hollows and large ponds relative to MOS and STE bogs (Fig. 1.1).

Lateral expansion of the peatlands is shown to have been an important factor on C accumulation. Although net vertical accumulation rates at the ecosystem scale have

diminished during the entire Holocene, maximum C accumulation rates were attained as late as the mid-Holocene, as basin geomorphological constraints apparently induced a delay in the culmination of ecosystem C sequestration rates in the three peatlands. Lateral expansion of peatlands has been reported to be primarily influenced by local factors as topography (Bauer, Gignac and Vitt, 2003 ; Korhola, 1994 ; Mäkilä and Moisanen, 2007). Therefore, site-typical patterns of lateral expansion may cause different temporal patterns of total C accumulation between peatlands. The local topography near LLC bog is complex (Fig. 1.2), with a steep outcrop at its western limit and a neighbouring stream in the eastern part. Hence, lateral peatland expansion may have been inhibited during development and this confinement could partially explain the decline in C accumulation. Belyea and Clymo (2001) showed that, in peatlands that are constrained in their development by mineral basin topography, the potential for water storage in the catotelm and thus peat accumulation decreases over time.

In addition to the spatial complexity of the studied peatlands, a nonlinear feedback to either allogenic or autogenic change may be present. The ecosystem's reaction to environmental change depends not only on the strength of the forcing but also on the height of the threshold value (Belyea, 2009). In this context, the resilience of peatland hydrology and vegetation might be of major importance when evaluating the reaction of the ecosystem to environmental change. As LLC bog surface topography is sloping uniformly, while MOS and STE bogs have more lens-shaped cross-sections, surface runoff patterns are likely to differ. Differential resilience between the Eastmain region peatlands might be another explanation for asynchronous shifts in C accumulation.

1.5.3. Allogenic factors

The gradual slowdown of net long-term C accumulation in LLC bog from 5000 to 250 cal BP and from 3500 to 250 cal BP in MOS and STE bogs (Fig. 1.7) corresponds to declining C sequestration rates observed in other continental peatlands in North America (Beaulieu-Audy *et al.*, 2009 ; Vardy *et al.*, 2000 ; Vitt *et al.*, 2000 ; Yu, 2006 ; Zoltai, 1995). This common trend may indicate the presence of an external factor such as climate change that could have mediated autogenic development. Holocene temperature reconstructions

show the presence of cooling starting between 4000 and 3000 cal BP for both the entire North American continent and northern Quebec (Viau *et al.*, 2006). Beaulieu-Audy *et al.* (2009) found slowdown of peat accumulation in two ombrotrophic peatlands north of the Eastmain region from 4500 to 1500 cal BP, whereas peatland permafrost aggradation was reported starting around 3500 cal BP (Bhiry, Payette and Robert, 2007) in the forest-tundra biome. A colder climate has also been responsible for a decline in forest productivity after 4650 cal BP about 150 km north of this study region (Arseneault and Sirois, 2004). Lower primary productivity caused by a prolonged period of peatland surface frost and shorter growing seasons may have suppressed C accumulation rates (*sensu* Mauquoy *et al.*, 2002). Slowdown of C sequestration in fens and bogs has also been reported from continental western Canada starting between 4000 and 3000 cal BP, in some cases linked to permafrost development (Vardy *et al.*, 2000 ; Vitt *et al.*, 2000 ; Yu, 2006 ; Zoltai, 1995).

As the Eastmain peatlands are at maximum 53 km apart, a spatially uniform climate regime can be assumed. The observed differences in timing and intensity of C accumulation trends between LLC bog and MOS and STE bogs seem inconsistent with the premise of a high sensitivity of peat C dynamics to Holocene hydroclimatic variations. Differences in past vegetation assemblages might explain different C sequestration patterns, yet analyses to this respect did not indicate a distinct difference in ombrotrophication between peatlands (Fig. 1.5). Vegetation shifts have been important throughout most of the peat cores, being as important as vegetation variability within and among peatlands (Fig. 1.5), which might imply a strong local hydrological control on vegetation.

Besides, isostatic uplift after the retreat of the Wisconsian ice sheet has been extremely high at around 220 m in the present-day James Bay region (Andrews and Peltier, 1989). As the total uplift shows a descending northwest-southeast gradient and the general drainage in the Eastmain region is northwestward, this differential uplift might have resulted in a decreasing slope of the Eastmain watershed region. Due to differential uplift, LLC bog may have risen 10 to 20 m more than MOS and STE bogs since 7000 cal BP. Such a difference may have influenced regional drainage through river incision. If incising rivers form a base level for peatland hydrology, this may result in a drawdown of the regional peatland water table mound. This mechanism has been proposed to explain variations in patterns of peatland development on the low-relief, southwestern Hudson Bay lowlands

(Glaser *et al.*, 2004). The differential uplift within the region studied by Glaser *et al.* (2004) is on the order of 40 m over the last 7000 years, with highest uplift occurring in the northeastern section. If the same dynamics were applicable in the Eastmain region, LLC bog, which could have suffered most from increases in regional drainage potential, would have registered most important bog surface drying and hence a slowing down of C accumulation. However, we do not have evidence of river incision at the eastern limit of LLC bog.

Stratigraphic analyses showed the presence of macroscopic charcoal fragments, indicating possible local burning events. Recurrent fires have a potential to affect long-term rates of accumulation in boreal peatlands, with emissions estimated at 2.2 to 2.5 kg C m^{-2} per event (Pitkänen, Turunen and Tolonen, 1999 ; Turetsky and Wieder, 2001). Although forest fires may not frequently affect wet, open peatland ecosystems (Hellberg, Niklasson and Granström, 2004), recent observations in the Eastmain region showed that fires may well burn central sections dominated by wet hollows after extreme drought. As the LLC bog is bordered by a steep ridge in the western part and the regional dominant fire direction is northwest-southeast (Bergeron *et al.*, 2004), LLC bog might have been less exposed to frequent fire events than MOS and STE bogs. However, as macroscopic charcoal fragments may easily be transported over several hundreds of meters (Peters and Higuera, 2007), we can not yet provide estimates of past frequencies of peatland burning in the Eastmain region. Peatland fire history linked to long-term C dynamics will be reconstructed for a future publication.

1.5.4. C accumulation rates

The mean basin C sequestration rate of the three peatlands is 16.2 g m^{-2} yr^{-1}. One should keep in mind that these rates are apparent and thus differ from net ecosystem production as millennia of deep decomposition have passed. This value is on the lower side of our hypothesized range of 15-25 g m^{-2} yr^{-1} that was based uniquely on present-day climate conditions (MAT and MAP). Thus, the difference between the hypothesized and the reconstructed values may be explained by differences in past climate regimes, climate variables other than MAT and MAP, disturbance regimes and geological (tectonic) and

geomorphological (substrate) factors. In eastern Canada, late-Holocene climate regimes are likely to have been colder than present-day, possibly explaining the relatively low C sequestration values. In addition, peatland development was delayed by late deglaciation. Hence, a relatively large part of the total Eastmain peatland C stock was sequestered during the less favourable Neoglacial conditions, resulting in suppressed mean Holocene C sequestration rates. These C accumulation rates are lower than the reported global northern averages of 24.1 g m^{-2} yr^{-1} (Lavoie, Paré and Bergeron, 2005) and 18.6 g m^{-2} yr^{-1} (Yu, Beilman and Jones, 2009). However, averages from single cores are likely to overestimate rates at the ecosystem scale as presented here; therefore direct comparisons may be hazardous. The obtained mean recent C accumulation rate of 56.4 g m^{-2} yr^{-1}, representing the recentmost 306 years of accumulation, is comparable to the mean ~150-year accumulation rate of eastern Canadian bogs of 73 g m^{-2} yr^{-1} (Turunen *et al.*, 2004) and 74 and 84 g m^{-2} yr^{-1} for other cores from LLC and MOS bogs (Loisel and Garneau, 2010). The lower values obtained in this study may be the result of the longer period considered for recent C accumulation rate calculation, as these values generally diminish with depth due to a higher proportion of decomposed peat (Turunen *et al.*, 2004).

At the global scale, highest long-term C accumulation rates may coincide with climates with MAT of 0-2.5°C and MAP of 400-550 mm (Yu, Beilman and Jones, 2009). The Eastmain region peatlands are close to the wet/cold limit of northern peatlands within the northern peatland distribution of Yu, Beilman and Jones (2009). To correctly interpret differing C sequestration patterns in relation to climate, the seasonal precipitation distribution and temperature patterns should be taken into account. The effect of precipitation on boreal peatland hydrology may vary depending on the precipitation type (i.e. rain or snow) (Charman, 2007), whereas warm summers but cold winters are likely to be favourable to C sequestration (Jones and Yu, 2010). In addition, it should be noted that the link between climate regime and temporal shifts in C sequestration is nonlinear. Gorham *et al.* (2003) stated that high rates of peat C accumulation are found in North American and Siberian dry continental regions. However, despite the good correlation between relatively dry climate and rapidly accumulating peatlands, dry shifts in climate will not in all cases cause increasing C accumulation rates. The reaction of a peatland to environmental change is mediated by microtopographic dynamics (e.g. bistability and spatial self-organization) (Eppinga *et al.*,

2009) showing an underestimated complexity of these systems. Thus, the direction of the ecosystem's pathway may depend on the present phytoecological and hydrological state relative to the climate shift, the rapidity of the forcing and the height of the threshold.

1.5.5. C density and mass per unit area

The mean C density from the studied peatlands of 44 kg m^{-3} is close to estimates from other regions. Vitt *et al.* (2000) obtained variable C densities depending on peat type: 49 kg m^{-3} for open fens and bogs and 55 kg m^{-3} for wooded and shrubby fens. Beilman *et al.* (2008) reported a mean of 93 kg OM m^{-3} in the boreal Mackenzie River Basin of northwestern Canada, which represents approximately 47 kg C m^{-3}. In Alaskan bogs with C sequestration rates similar to those of the Eastmain peatlands, mean C densities of 42-47 kg m^{-3} were obtained compared to 35-37 kg C m^{-3} south of the Eastmain region (Gorham, Janssens and Glaser, 2003). In western Siberia, a mean ombrotrophic peat C density of 32 kg m^{-3} was reported (Bleuten and Lapshina, 2001). Assuming 50% of C in OM, 46 kg m^{-3} was obtained by Turunen *et al.* (2001) for west-Siberian peatlands and a mean C density of 36 kg m^{-3} was found in a hummock-hollow pine bog in southeastern Finland (Mäkilä, 1997). Important differences between peatlands within a region may indicate that local and short-term factors as vegetation type and hydrology may be determinant considering C density values.

Mean area-weighted C mass per unit area for the Eastmain peatlands is 91 kg m^{-2}. This value is lower than 131 kg m^{-2}, 118 kg m^{-2} and 119 kg m^{-2} obtained from peatlands in (south)western Canada (Vitt *et al.*, 2000), northwestern Canada (Beilman *et al.*, 2008) and the western Siberian lowlands (Sheng *et al.*, 2004), respectively. It is also lower than the mean 124 kg m^{-2} and 108 kg m^{-2} for the raised bog and aapa mire region in Finland (Mäkilä and Goslar, 2008) and 94 kg m^{-2} as an average for Scottish peatlands (Chapman *et al.*, 2009). Comparisons between regions are hampered by methodological differences, which partially explain lower values for Eastmain region peatlands.

1.5.6. Future perspective on potential C sequestration

To estimate future C sequestration patterns in the Eastmain region, a detailed image of the factors driving long-term C accumulation rate decline is crucial. For now, we do not have indications that either autogenic or allogenic factors are dominant due to the complexity of these ecosystems. In case the decline can be attributed primarily to autogenic factors as a height-induced long-term drying of the surface (*sensu* Yu *et al.*, 2003), the C sink of the Eastmain region may continue to diminish during the centuries to come. However, if long-term Neoglacial cooling has been the principal cause, the projected important warming trend with higher precipitation in the Eastmain region (Plummer *et al.*, 2006) might reverse the trend and increase the potential for C accumulation as registered from contemporaneous measurements between 2006-2008 (Pelletier, Garneau and Moore, in review). Considering C sequestration, there is probably a close imbrication of both autogenic and allogenic factors on various timescales that requires better understanding.

1.6. Conclusion

This paper presents the first Holocene C accumulation rates from northern Quebec boreal bogs, with a mean value of 16.2 g m^{-2} yr^{-1}. Ecosystem C flux reconstructions for the Eastmain region show declining values at the onset of the late-Holocene. This slowdown of C accumulation is principally the result of a decrease in net rate of vertical accumulation. The application of the age-depth model to the entire ecosystem results in a model for lateral expansion, showing that the increase in peatland area was probably important during the early development and that ecosystem C fluxes became high once extensive area was covered by peat between 5000 and 3000 cal BP.

Variation in timing of the onset of the decline in C accumulation rates and the intensity of the slowdown between peatlands shows the influence of site-specific factors and local ecosystem complexity, however, a less accurate age-depth model for MOS and STE bogs may have caused a bias for the early stages of accumulation. Long-term diminishing rates of C accumulation in North American peatlands may be associated with two

phenomena. First, potential for peat accumulation is suspected to decrease with time in peatlands that are constrained in lateral expansion (Belyea and Clymo, 2001). Second, late-Holocene cooling may well have caused changes in C sequestration rates. Changes in peatland vegetation, hydrology and fire regimes as well as the formation of permafrost since ~4000 cal BP have been reported at the continent scale (e.g. Beaulieu-Audy *et al.*, 2009 ; Bhiry, Payette and Robert, 2007 ; Vitt *et al.*, 2000 ; Zoltai, 1995).

Lateral peatland expansion is shown to be an important factor considering C sequestration at the ecosystem scale and comparisons between cores show the complexity of peatland ecosystems. Thus, in order to obtain more accurate quantifications of past C accumulation and better understand the role of peatlands in the global C cycle, multiple cores should be considered, rather than a unique central core. To estimate the direction of C accumulation trends in the Eastmain region for the centuries to come, the principal driving mechanism needs to be identified, as the principal autogenic and allogenic factors may have a contrasting influence on long-term C accumulation.

Acknowledgments

We thank Hans Asnong, Maxime Boivin, Gabrièle Guay, Claire Lacroix, Sébastien Lacoste, Julie Loisel, Éric Rosa and Charles Vaillancourt for field assistance. Thanks to Pierre J.H. Richard (Université de Montréal) for the use of the Jacques-Rousseau laboratory for analyses. We greatly appreciate the assistance by Bertrand Fournier (UQAM) with statistical analyses. Funding was provided by Hydro-Quebec Production through the EM-1 Project *Reservoirs' net greenhouse gas emissions research project (2005-2009)*. We thank the hydroclimatological scenarios team of the Ouranos consortium for providing climate variables by manipulation of 1971-2003 NLWIS data. We appreciate the useful comments of two anonymous reviewers. Thanks to *Les Tourbeux* for discussions and inspiration.

References

Andrews, J. A., and W. R. Peltier. 1989. «Quaternary geodynamics in Canada». In *Quaternary geology of Canada and Greenland,* R. J. Fulton, p. 543-572: Geological Survey of Canada, no. 1.

Arseneault, D., and L. Sirois. 2004. «The millennial dynamics of a boreal forest stand from buried trees». *Journal of Ecology*. vol. 92, p. 490-504.

Bauer, I. E., J. S. Bhatti, C. Swanston, R. K. Wieder and C. M. Preston. 2009. «Organic matter accumulation and community change at the peatland-upland interface: Inferences from ^{14}C and ^{210}Pb dated profiles». *Ecosystems*. vol. 12, no 4, p. 636-653.

Bauer, I. E., L. D. Gignac and D. H. Vitt. 2003. «Development of a peatland complex in boreal western Canada: Lateral site expansion and local variability in vegetation succession and long-term peat accumulation». *Canadian Journal of Botany*. vol. 81, no 8, p. 833-847.

Beaulieu-Audy, V., M. Garneau, P. J. H. Richard and H. Asnong. 2009. «Holocene palaeoecological reconstruction of three boreal peatlands in the La Grande Riviere region, Quebec, Canada». *The Holocene*. vol. 19, no 3, p. 459-476.

Beilman, D. W., G. M. MacDonald, L. C. Smith and P. J. Reimer. 2009. «Carbon accumulation in peatlands of West Siberia over the last 2000 years». *Global Biogeochemical Cycles*. vol. 23, no GB1012, p. doi:10.1029/2007GB003112.

Beilman, D. W., D. H. Vitt, J. S. Bhatti and S. Forest. 2008. «Peat carbon stocks in the southern Mackenzie River Basin: uncertainties revealed in a high-resolution case study». *Global Change Biology*. vol. 14, no 6, p. 1221-1232.

Belyea, L. R. 2009. «Nonlinear dynamics of peatlands and potential feedbacks on the climate system». In *Carbon cycling in northern peatlands,* Andrew J. Baird, Lisa R. Belyea, Xavier Comas, A. S. Reeve et Lee D. Slater, p. 5-18. Washington: American Geophysical Union.

Belyea, L. R., and A. J. Baird. 2006. «Beyond "the limits to peat bog growth": cross-scale feedback in peatland development». *Ecological Monographs*. vol. 76, no 3, p. 299-322.

Belyea, L. R., and R. S. Clymo. 2001. «Feedback control of the rate of peat formation». *Proceedings: Biological Sciences*. vol. 268, no 1473, p. 1315-1321.

Belyea, L. R., and N. Malmer. 2004. «Carbon sequestration in peatland: patterns and mechanisms of response to climate change». *Global Change Biology*. vol. 10, no 7, p. 1043-1052.

Bergeron, Y., S. Gauthier, M. Flannigan and V. Kafka. 2004. «Fire regimes at the transition between mixedwood and coniferous boreal forest in northwestern Quebec». *Ecology*. vol. 85, no 7, p. 1916-1932.

Bhiry, N., S. Payette and É. C. Robert. 2007. «Peatland development at the arctic tree line (Québec, Canada) influenced by flooding and permafrost». *Quaternary Research*. vol. 67, no 3, p. 426-437.

Blaauw, M., and J. A. Christen. 2005. «Radiocarbon peat chronologies and environmental change». *Journal of the Royal Statistical Society. Series C: Applied Statistics* vol. 54, no 4, p. 805-816.

Bleuten, W., and E. D. Lapshina. 2001. *Carbon storage and atmospheric exchange by west Siberian peatlands*. Utrecht: Physical Geography, Universiteit Utrecht, 169 p.

Borren, W., and W. Bleuten. 2006. «Simulating Holocene carbon accumulation in a western Siberian watershed mire using a three-dimensional dynamic modeling approach». *Water Resources Research*. vol. 42, no W12413, p. doi:10.1029/2006WR004885.

Borren, W., W. Bleuten and E. D. Lapshina. 2004. «Holocene peat and carbon accumulation rates in the southern taiga of western Siberia». *Quaternary Research*. vol. 61, no 1, p. 42-51.

Bubier, J. L. 1991. «Patterns of Picea mariana (black spruce) growth and raised bog development in Victory Basin, Vermont». *Bulletin of the Torrey Botanical Club*. vol. 118, no 4, p. 399-411.

Bubier, J. L., P. Crill, A. Mosedale, S. Frolking and E. Linder. 2003. «Peatland responses to varying interannual moisture conditions as measured by automatic CO_2 chambers». *Global Biogeochemical Cycles*. vol. 17, no 2, 1066, p. doi:10.1029/2002GB001946.

Chapman, S. J., J. Bell, D. Donnelly and A. Lilly. 2009. «Carbon stocks in Scottish peatlands». *Soil Use and Management*. vol. 25, no 2, p. 105-112.

Charman, D. J. 2002. *Peatlands and environmental change*. Chichester: John Wiley & Sons Ltd., 301 p.

Charman, D. J. 2007. «Summer water deficit variability controls on peatland water-table changes: implications for Holocene palaeoclimate reconstructions». *The Holocene*. vol. 17, no 2, p. 217-227.

Clymo, R. S. 1984. «The limits to peat bog growth». *Philosophical Transactions of the Royal Society of London. Series B, Biological Sciences*. vol. 303, no 1117, p. 605-654.

Clymo, R. S., J. Turunen and K. Tolonen. 1998. «Carbon accumulation in peatland». *Oikos*. vol. 81, no 2, p. 368-388.

Dean, W. E. 1974. «Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition; comparison with other methods». *Journal of Sedimentary Research*. vol. 44, no 1, p. 242-248.

Dorrepaal, E., S. Toet, R. S. P. van Logtestijn, E. Swart, M. J. van de Weg, T. V. Callaghan and R. Aerts. 2009. «Carbon respiration from subsurface peat accelerated by climate warming in the subarctic». *Nature*. vol. 460, no 7255, p. 616-619.

Dyke, A. S., A. Moore and L. Robertson (2003). Deglaciation of North America. Ottawa, Geological Survey of Canada, Natural Resources Canada

Eppinga, M. B., M. Rietkerk, M. J. Wassen and P. C. De Ruiter. 2009. «Linking habitat modification to catastrophic shifts and vegetation patterns in bogs». *Plant Ecology*. vol. 200, p. 53-68.

Frolking, S., and N. T. Roulet. 2007. «Holocene radiative forcing impact of northern peatland carbon accumulation and methane emissions». *Global Change Biology*. vol. 13, no 5, p. 1079-1088.

Glaser, P. H., B. Hansen, C. S., D. I. Siegel, A. S. Reeve and P. J. Morin. 2004. «Rates, pathways and drivers for peatland development in the Hudson Bay Lowlands, northern Ontario, Canada». *Journal of Ecology*. vol. 92, no 6, p. 1036-1053.

Gorham, E. 1991. «Northern peatlands: role in the carbon cycle and probable responses to climatic warming». *Ecological Applications*. vol. 1, no 2, p. 182-195.

Gorham, E., J. A. Janssens and P. H. Glaser. 2003. «Rates of peat accumulation during the postglacial period in 32 sites from Alaska to Newfoundland, with special emphasis on northern Minnesota». *Canadian Journal of Botany*. vol. 81, no 5, p. 429-438.

Grenier, M., S. Labrecque, M. Garneau and A. Tremblay. 2008. «Object-based classification of a SPOT-4 image for mapping wetlands in the context of greenhouse gases emissions: the case of the Eastmain region, Québec, Canada». *Canadian Journal of Remote Sensing*. vol. 34, no 2, p. S398-S314.

Hänninen, P. 1992. «Application of Ground Penetrating Radar and radio wave moisture probe techniques to peatland investigations». Rovaniemi, Geological Survey of Finland, 71 p.

Haslett, J., and A. Parnell. 2008. «A simple monotone process with application to radiocarbon-dated depth chronologies». *Journal of the Royal Statistical Society: Series C*. vol. 57, no 4, p. 399-418.

Heiri, O., A. F. Lotter and G. Lemcke. 2001. «Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results». *Journal of Paleolimnology*. vol. 25, no 1, p. 101-110.

Hellberg, E., M. Niklasson and A. Granström. 2004. «Influence of landscape structure on patterns of forest fires in boreal forest landscapes in Sweden». *Canadian Journal of Forest Research*. vol. 34, p. 332-338.

Hutchinson, M. F., D. W. McKenney, K. Lawrence, J. H. Pedlar, R. F. Hopkinson, E. Milewska and P. Papadopol. 2009. «Development and testing of Canada-wide interpolated spatial models of daily minimum–maximum temperature and precipitation for 1961–2003». *Journal of Applied Meteorology and Climatology*. vol. 48, p. 725-741.

Jol, H. M., and C. S. Bristow. 2003. «GPR in sediments: advice on data collection, basic processing and interpretation, a good practice guide». *Geological Society, London, Special Publications*. vol. 211, no 1, p. 9-27.

Jones, M. C., and Z. Yu. 2010. «Rapid deglacial and early Holocene expansion of peatlands in Alaska». *Proceedings of the National Academy of Sciences*. vol. 107, no 16, p. 7347-7352.

Korhola, A. 1994. «Radiocarbon evidence for rates of lateral expansion in raised mires in southern Finland». *Quaternary Research*. vol. 42, no 3, p. 299-307.

Korhola, A., J. Alm, K. Tolonen, J. Turunen and H. Jungner. 1996. «Three-dimensional reconstruction of carbon accumulation and CH_4 emission during nine millennia in a raised mire». *Journal of Quaternary Science*. vol. 11, no 2, p. 161-165.

Korhola, A., M. Ruppel, H. Seppä, M. Väliranta, T. Virtanen and J. Weckström. 2010. «The importance of northern peatland expansion to the late-Holocene rise of atmospheric methane». *Quaternary Science Reviews*. vol. 29, no 5-6, p. 611-617.

Kuhry, P., and D. H. Vitt. 1996. «Fossil carbon/nitrogen ratios as a measure of peat decomposition». *Ecology*. vol. 77, no 1, p. 271-275.

Lavoie, M., D. Paré and Y. Bergeron. 2005. «Impact of global change and forest management on carbon sequestration in northern forested peatlands». *Environmental Reviews*. vol. 13, p. 199-240.

Leopold, M., and J. Volkel. 2003. «GPR images of periglacial slope deposits beneath peat bogs in the Central European Highlands, Germany». *Geological Society, London, Special Publications*. vol. 211, no 1, p. 181-189.

Loisel, J., and M. Garneau. 2010. «Late Holocene paleoecohydrology and carbon accumulation estimates from two boreal peat bogs in eastern Canada: Potential and limits of multi-proxy archives». *Palaeogeography, Palaeoclimatology, Palaeoecology*. vol. 291, no 3-4, p. 493-533.

MacDonald, G. M., D. W. Beilman, K. V. Kremenetski, Y. Sheng, L. C. Smith and A. A. Velichko. 2006. «Rapid early development of circumarctic peatlands and atmospheric CH_4 and CO_2 variations». *Science*. vol. 314, no 5797, p. 285-288.

Mäkilä, M. 1997. «Holocene lateral expansion, peat growth and carbon accumulation on Haukkasuo, a raised bog in southeastern Finland». *Boreas*. vol. 26, no 1, p. 1-14.

Mäkilä, M., and T. Goslar. 2008. «The carbon dynamics of surface peat layers in southern and central boreal mires of Finland and Russian Karelia». *Suo*. vol. 59, no 3, p. 49-69.

Mäkilä, M., and M. Moisanen. 2007. «Holocene lateral expansion and carbon accumulation of Luovuoma, a northern fen in finnish Lapland». *Boreas*. vol. 36, no 2, p. 198-210.

Malmer, N., and B. Wallén. 1999. «The dynamics of peat accumulation on bogs: mass balance of hummocks and hollows and its variation throughout a millennium». *Ecography*. vol. 22, no 6, p. 736-750.

Mauquoy, D., T. Engelkes, M. H. M. Groot, F. Markesteijn, M. G. Oudejans, J. van der Plicht and B. van Geel. 2002. «High-resolution records of late-Holocene climate change and carbon accumulation in two north-west European ombrotrophic peat bogs». *Palaeogeography, Palaeoclimatology, Palaeoecology*. vol. 186, no 3-4, p. 275-310.

Neal, A. 2004. «Ground-penetrating radar and its use in sedimentology: principles, problems and progress». *Earth-Science Reviews*. vol. 66, no 3-4, p. 261-330.

Nilsson, M., M. Klarqvist, E. Bohlin and G. Possnert. 2001. «Variation in ^{14}C age of macrofossils and different fractions of minute peat samples dated by AMS». *The Holocene*. vol. 11, no 5, p. 579-586.

Nordbakken, J.-F. 1996. «Plant niches along the water-table gradient on an ombrotrophic mire expanse». *Ecography*. vol. 19, no 2, p. 114-121.

Payette, S. 1988. «Late-Holocene development of subarctic ombrotrophic peatlands: allogenic and autogenic succession». *Ecology*. vol. 69, no 2, p. 516-531.

Payette, S., and L. Rochefort. 2001. *Écologie des tourbières du Québec-Labrador* Trad. de: *French*. Ste-Foy: Les presses de l'Université Laval, 621 p.

Pelletier, L., M. Garneau and T. R. Moore. in review. «Interannual variation in net ecosystem exchange at the microform scale in a boreal bog, Eastmain region, Quebec, Canada». *Journal of Geophysical Research - Biogeosciences*.

Peters, M. E., and P. E. Higuera. 2007. «Quantifying the source area of macroscopic charcoal with a particle dispersal model». *Quaternary Research*. vol. 67, no 2, p. 304-310.

Pitkänen, A., J. Turunen and K. Tolonen. 1999. «The role of fire in the carbon dynamics of a mire, eastern Finland». *The Holocene*. vol. 9, no 4, p. 453-462.

Plummer, D. A., D. Caya, A. Frigon, H. Côté, M. Giguère, D. Paquin, S. Biner, R. Harvey and R. de Elia. 2006. «Climate and climate change over North America as simulated by the Canadian RCM». *Journal of Climate*. vol. 19, no 13, p. 3112-3132.

Reimer, P. J., M. G. L. Baillie, E. Bard, A. Bayliss, J. W. Beck, C. J. H. Bertrand, P. G. Blackwell, C. E. Buck, G. S. Burr, K. B. Cutler, P. E. Damon, R. L. Edwards, R. G. Fairbanks, M. Friedrich, T. P. Guilderson, A. G. Hogg, K. A. Hughen, B. Kromer, G. McCormac, S. Manning, C. Bronk Ramsey, R. W. Reimer, S. Remmele, J. R.

Southon, M. Stuiver, S. Talamo, F. W. Taylor, J. van der Plicht and C. E. Weyhenmeyer. 2004. «INTCAL04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP». *Radiocarbon*. vol. 46, no 3, p. 1029-1058.

Rosa, E., M. Larocque, S. Pellerin, S. Gagné and B. Fournier. 2009. «Determining the number of manual measurements required to improve peat thickness estimations by ground penetrating radar». *Earth Surface Processes and Landforms*. vol. 34, no 3, p. 377-383.

Roulet, N. T. 2000. «Peatlands, carbon storage, greenhouse gases, and the Kyoto protocol: prospects and significance for Canada». *Wetlands*. vol. 20, no 4, p. 605-615.

Rydin, H. 1993. «Interspecific competition between Sphagnum mosses on a raised bog». *Oikos*. vol. 66, no 3, p. 413-423.

Sandmeier, K. J. (2005). Reflexw 3.5.7. Manual. Karlsruhe, Sandmeier Software

Saucier, J.-P., J.-F. Bergeron, P. Grondin and A. Robitaille. 1998. «Les régions écologiques du Québec méridional (3e version): un des éléments du système hiérarchique de classification écologique du territoire mis au point par le ministère des Ressources naturelles du Québec». *L'Aubelle*. vol. 124, p. 1-12.

Schimel, D. S., J. I. House, K. A. Hibbard, P. Bousquet, P. Ciais, P. Peylin, B. H. Braswell, M. J. Apps, D. Baker, A. Bondeau, J. Canadell, G. Churkina, W. Cramer, A. S. Denning, C. B. Field, P. Friedlingstein, C. Goodale, M. Heimann, R. A. Houghton, J. M. Melillo, B. Moore III, D. Murdiyarso, I. Noble, S. W. Pacala, I. C. Prentice, M. R. Raupach, P. J. Rayner, R. J. Scholes, W. L. Steffen and C. Wirth. 2001. «Recent patterns and mechanisms of carbon exchange by terrestrial ecosystems». *Nature*. vol. 414, p. 169-172.

Sheng, Y., L. C. Smith, G. M. MacDonald, K. V. Kremenetski, K. E. Frey, A. A. Velichko, M. Lee, D. W. Beilman and P. Dubinin. 2004. «A high-resolution GIS-based inventory of the west Siberian peat carbon pool». *Global Biogeochemical Cycles*. vol. 18.

Swanson, D. K. 2007. «Interaction of mire microtopography, water supply, and peat accumulation in boreal mires». *Suo*. vol. 58, no 2, p. 37-47.

Tolonen, K., and J. Turunen. 1996. «Accumulation rates of carbon in mires in Finland and implications for climate change». *The Holocene*. vol. 6, no 2, p. 171-178.

Troels-Smith, J. 1955. «Characterization of unconsolidated sediments ». *Danmarks Geologiske Undersogelse, Series IV*. vol. 3, p. 38-73.

Turetsky, M., K. Wieder, L. Halsey and D. Vitt. 2002. «Current disturbance and the diminishing peatland carbon sink». *Geophysical Research Letters*. vol. 29, no 11, 1526, p. doi:10.1029/2001GL014000.

Turetsky, M. R., and R. K. Wieder. 2001. «A direct approach to quantifying organic matter lost as a result of peatland wildfire». *Canadian Journal of Forest Research*. vol. 31, p. 363-366.

Turetsky, M. R., R. K. Wieder, D. H. Vitt, R. J. Evans and K. D. Scott. 2007. «The disappearance of relict permafrost in boreal north America: effects on peatland carbon storage and fluxes». *Global Change Biology*. vol. 13, p. 1922-1934.

Turunen, J., N. T. Roulet, T. R. Moore and P. J. H. Richard. 2004. «Nitrogen deposition and increased carbon accumulation in ombrotrophic peatlands in eastern Canada». *Global Biogeochemical Cycles*. vol. 18, no GB3002, p. doi:10.1029/2003GB002154.

Turunen, J., T. Tahvanainen, K. Tolonen and A. Pitkänen. 2001. «Carbon accumulation in West Siberian mires, Russia». *Global Biogeochemical Cycles*. vol. 15, no 2, p. 285-296.

Turunen, J., E. Tomppo, K. Tolonen and A. Reinikainen. 2002. «Estimating carbon accumulation rates of undrained mires in Finland-application to boreal and subarctic regions». *The Holocene*. vol. 12, no 1, p. 69-80.

Vardy, S. R., B. G. Warner, J. Turunen and R. Aravena. 2000. «Carbon accumulation in permafrost peatlands in the Northwest Territories and Nunavut, Canada». *The Holocene*. vol. 10, no 2, p. 273-280.

Viau, A. E., K. Gajewski, M. C. Sawada and P. Fines. 2006. «Millennial-scale temperature variations in North America during the Holocene». *Journal of Geophysical Research*. vol. 111.

Vitt, D. H., L. A. Halsey, I. E. Bauer and C. Campbell. 2000. «Spatial and temporal trends in carbon storage of peatlands of continental western Canada through the Holocene». *Canadian Journal of Earth Sciences*. vol. 37, no 5, p. 683-693.

Waddington, J. M., and N. T. Roulet. 1999. «Carbon balance of a boreal patterned peatland». *Global Change Biology*. vol. 6, p. 87-97.

Yu, Z. 2006. «Holocene carbon accumulation of fen peatlands in boreal western Canada: a complex ecosystem response to climate variation and disturbance». *Ecosystems*. vol. 9, no 8, p. 1278-1288.

Yu, Z., D. W. Beilman and M. C. Jones. 2009. «Sensitivity of northern peatland carbon dynamics to Holocene climate change». In *Carbon cycling in northern peatlands,* Andrew J. Baird, Lisa R. Belyea, Xavier Comas, A. S. Reeve et Lee D. Slater, p. 55-69. Washington: American Geophysical Union.

Yu, Z., J. Loisel, D. P. Brosseau, D. W. Beilman and S. J. Hunt. 2010. «Global peatland dynamics since the Last Glacial Maximum». *Geophysical Research Letters*. vol. 37, no 13, p. L13402.

Yu, Z., D. H. Vitt, I. D. Campbell and M. J. Apps. 2003. «Understanding Holocene peat accumulation pattern of continental fens in western Canada». *Canadian Journal of Botany*. vol. 81, p. 267-282.

Zoltai, S. C. 1995. «Permafrost distribution in peatlands of West-Central Canada during the Holocene Warm Period 6000 years BP». *Géographie physique et Quaternaire*. vol. 49, no 1, p. 45-54.

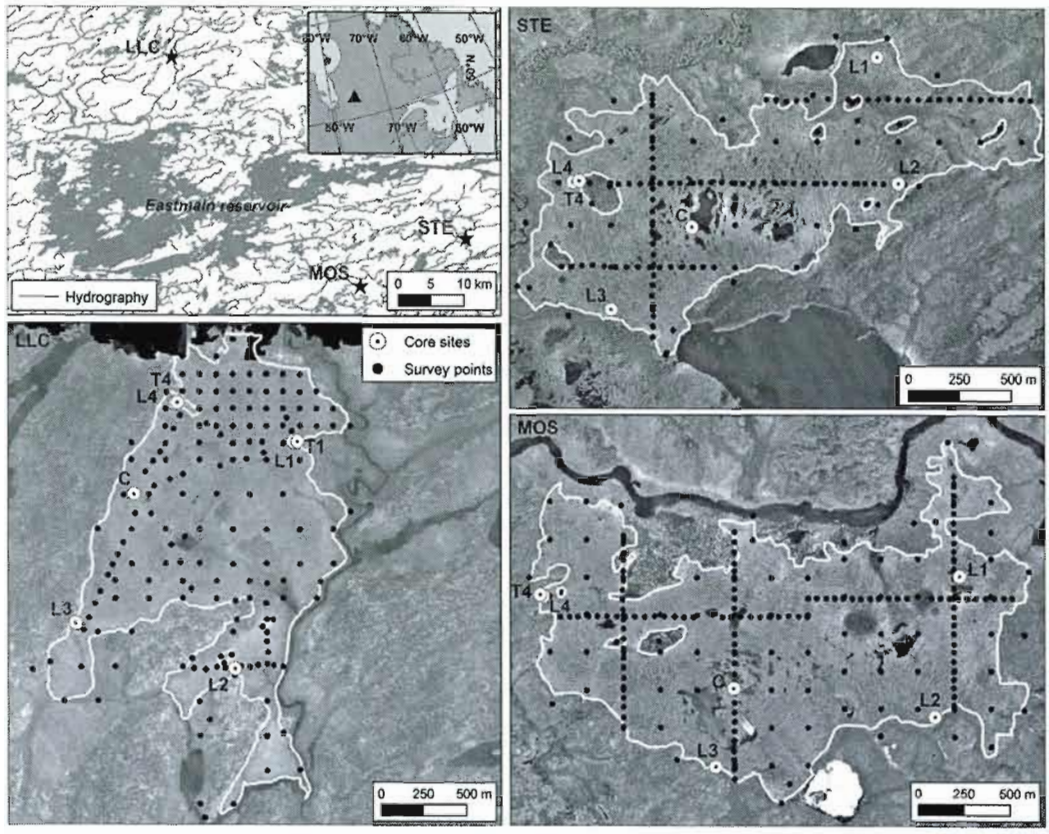

Figure 1.1. Location of the Eastmain region within eastern Canada and studied peatlands.

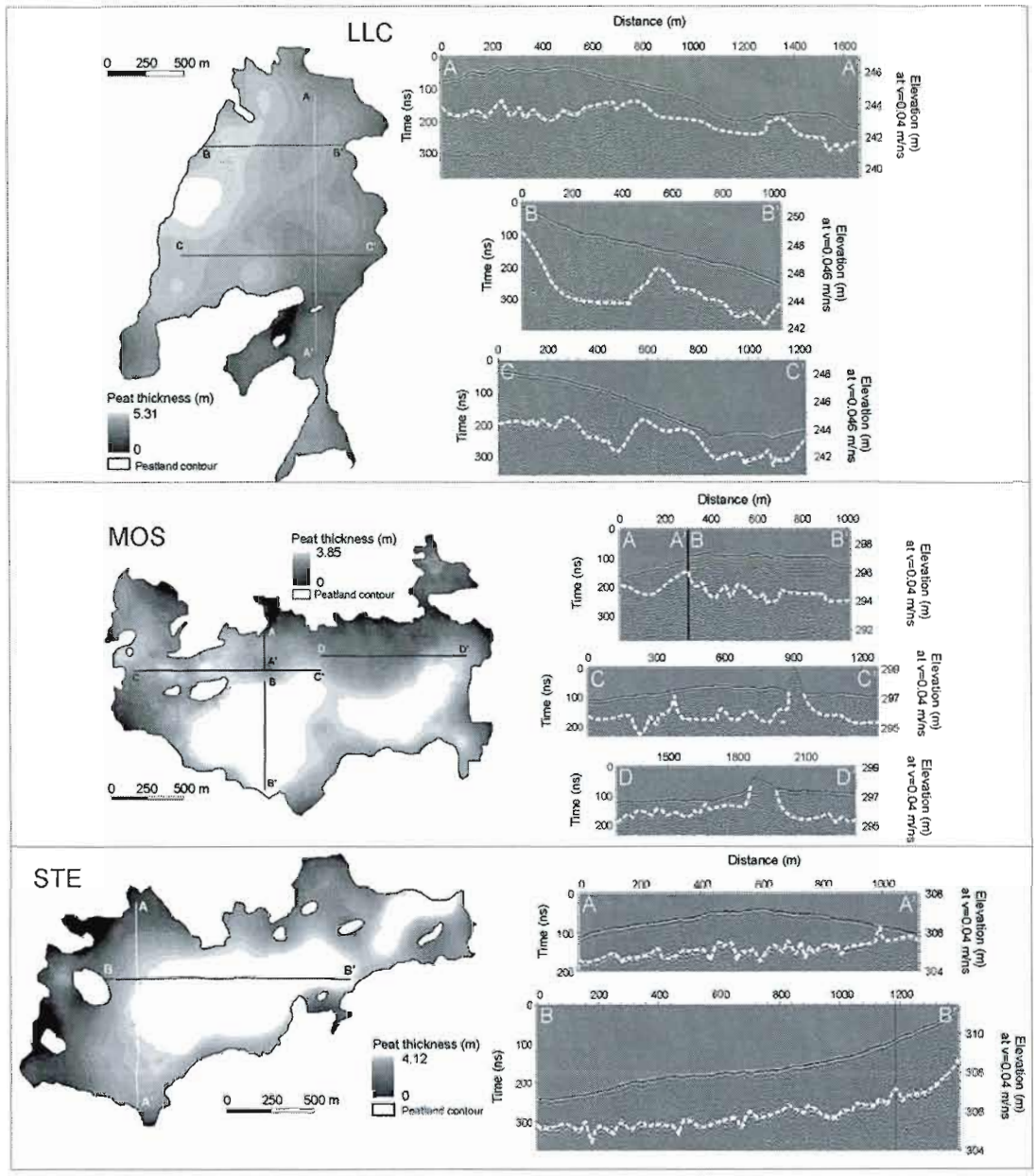

Figure 1.2. Peat cover thickness models and GPR profiles for each peatland. For better visibility, the mineral-organic contact is accentuated by a white dashed line in the profiles.

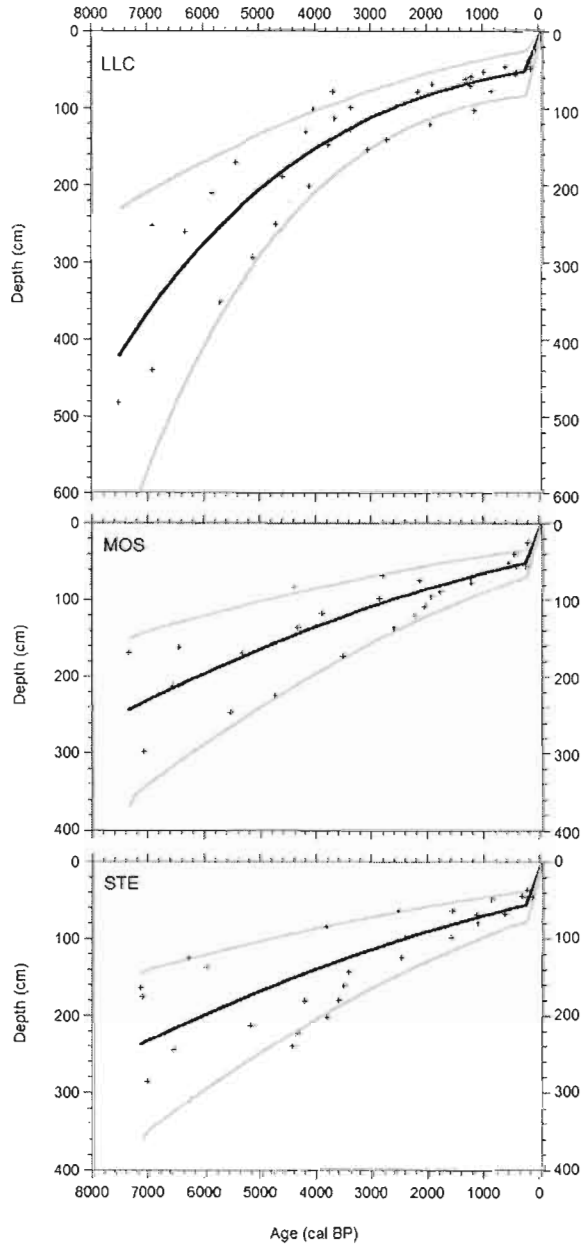

Figure 1.3. Age-depth models for each of the peatlands. Models were based on radiocarbon datings from 7 (LLC) or 6 (MOS and STE) cores per peatland. Grey lines indicate the limits of the 95% confidence interval.

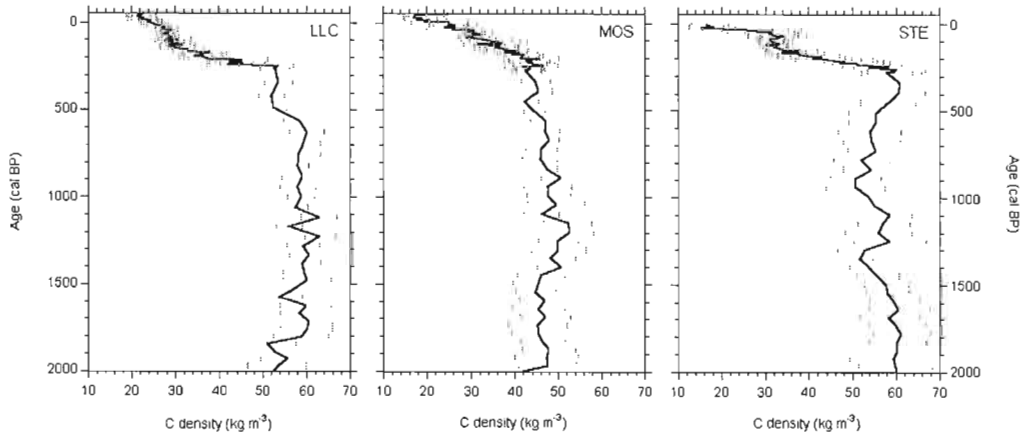

Figure 1.4. Mean C density since 2000 cal BP for each peatland. Brackets indicate the standard error. Means and standard errors were calculated from five cores per peatland.

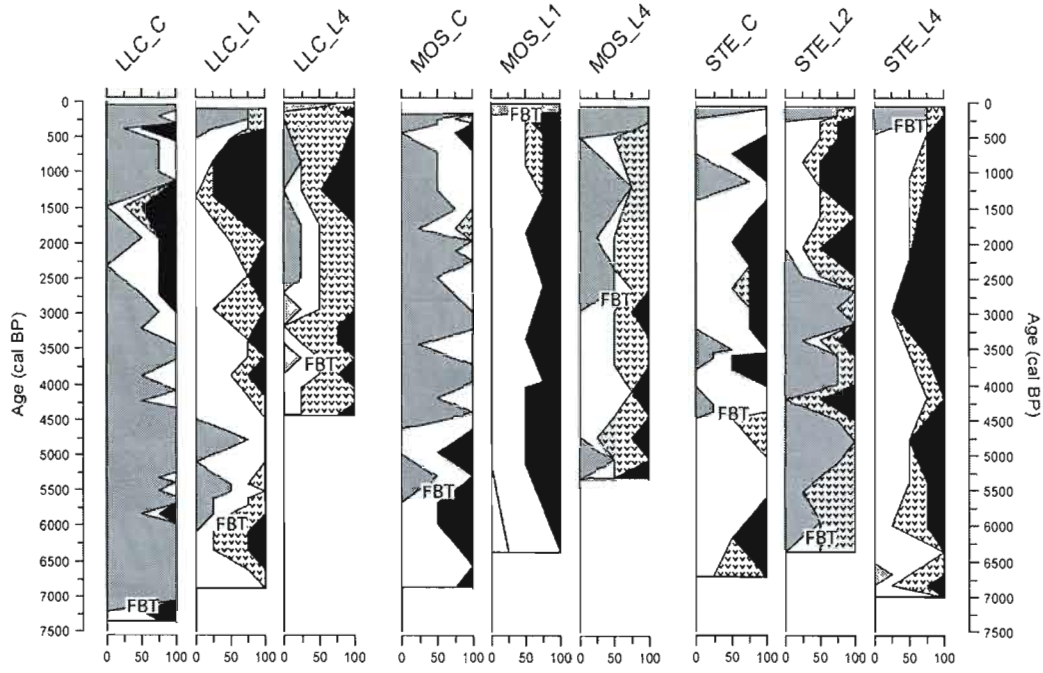

Figure 1.5. Peatland vegetation types for each peatland, expressed as volume percentages. Columns show *Sphagnum* peat (dark grey), moss (non-*Sphagnum*) peat (light grey), herbaceous peat (white), wood peat (v-symbol) and unidentifiable organic matter (black). Timing of the fen-bog transition is indicated by FBT for each core.

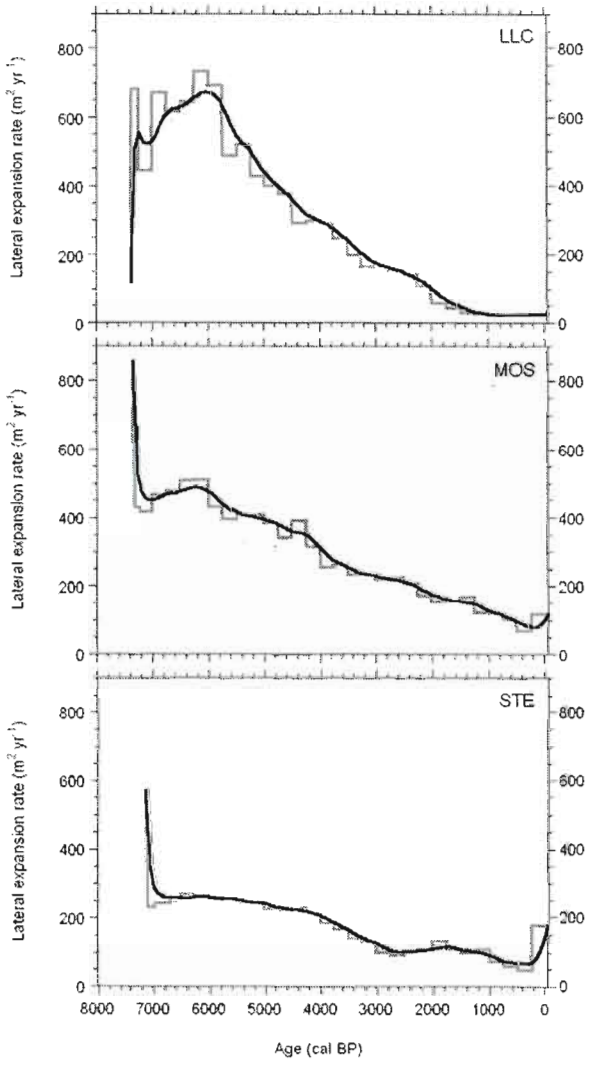

Figure 1.6. Peatland lateral expansion rates during the Holocene. Black line represents smoothing.

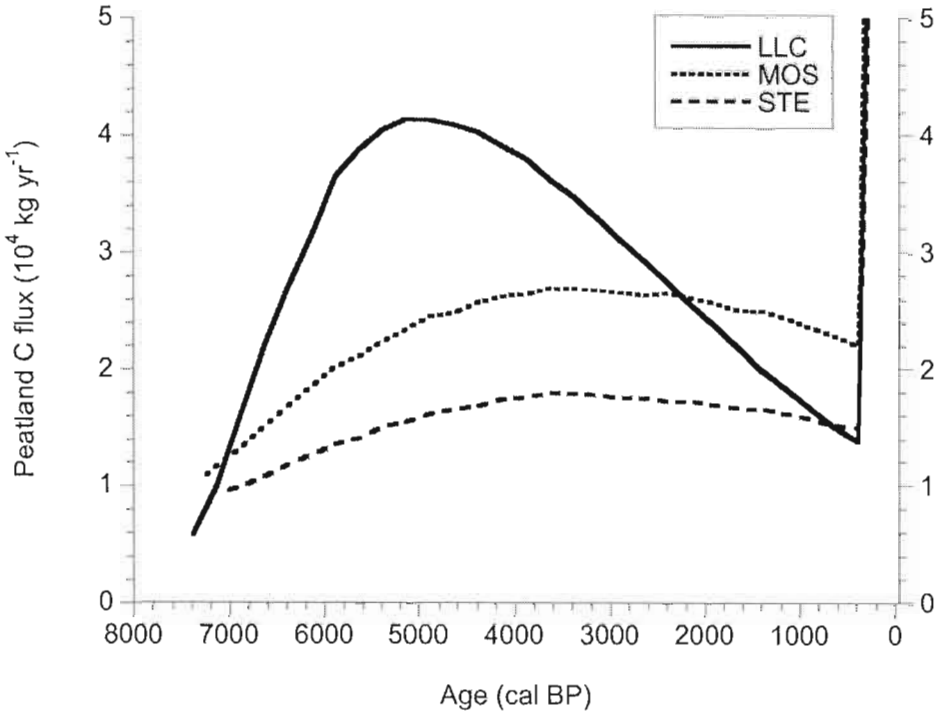

Figure 1.7. Holocene ecosystem C fluxes for each peatland.

Site	Core	Sample depth (cm)	Laboratory number	Material	^{14}C age (BP)	2σ range (cal BP)	Age (cal BP)
LLC	C	51-52	UCIAMS43480	Sph stems	340±20	317-480	391
	C	77-78	UCIAMS58634	Sph stems	915±15	777-906	846
	C	102-103	UCIAMS50203	Sph stems	1205±15	1072-1225	1140
	C	120-121	UCIAMS57419	Sph stems	1980±15	1872-1991	1924
	C	140-141	UCIAMS57421	Sph stems	2550±15	2531-2748	2716
	C	153-154	UCIAMS50204	Sph stems	2915±15	2984-3155	3057
	C	201-202	UCIAMS43479	Sph stems	3745±20	3996-4171	4107
	C	250-251	UCIAMS58636	Sph stems	4165±20	4596-4819	4701
	C	293-294	UCIAMS50205	Sph stems	4450±15	4980-5265	5110
	C	351-352	UCIAMS43478	Sph stems	4985±20	5653-5844	5701
	C	439-440	UCIAMS50206	Sph stems	6055±15	6821-6968	6912
	C	480-483	Beta223743	Eric leaf frs	6640±40	7431-7627	7520
	L1	45-46	UCIAMS58637	Sph stems	630±15	551-670	601
	L1	58-59	UCIAMS54956	Sph stems	1250±30	1078-1289	1207
	L1	68-69	UCIAMS58639	Sph stems, Larix/Picea leaf frs	1940±20	1810-1975	1889
	L1	99-100	UCIAMS57423	Sph stems	3125±15	3255-3390	3355
	L1	112-113	UCIAMS58638	Sph stems	3395±15	3586-3702	3656
	L1	130-131	UCIAMS54955	Sph stems	3780±25	4056-4269	4162
	L1	170-171	UCIAMS57418	Sph stems	4625±15	5295-5447	5418
	L1	210-211	UCIAMS64581	Sph stems	5035±20	5721-5903	5841
	L1	249-254	UCIAMS40365	Sph stems, Eric leaf frs	6055±20	6673-7218	6908
	L2	100-101	UCIAMS40366	Picea leaf frs	3690±20	3971-4098	4035
	L3	258-261	UCIAMS40367	Picea/Eric leaf frs	5500±20	6275-6383	6329
	L4	39-40	UCIAMS57417	Sph stems, Eric/Picea leaf frs	190±15	151-294	198
	L4	52-53	UCIAMS58632	Sph stems	1070±15	933-1075	978
	L4	61-63	UCIAMS58633	Sph stems, Eric leaf frs	1405±20	1216-1427	1301
	L4	78-79	UCIAMS57415	Sph stems, Larix/Eric leaf frs	2170±15	2129-2317	2152
	L4	127-128	UCIAMS57422	Sph stems	3135±15	3342-3435	3351
	L4	147-148	UCIAMS58635	Sph stems, Eric/Larix leaf frs	3495±20	3670-3826	3769
	L4	188-189	UCIAMS40368	Sph stems	4120±20	4520-4788	4586
	T1	47-48	UCIAMS64582	Sph stems	105±20	34-262	126
	T1	53-54	UCIAMS54957	Sph stems	300±25	304-691	410
	T1	77-80	UCIAMS54954	Sph stems	3440±25	3007-4270	3684
	T4	45-46	UCIAMS57416	Sph stems	180±15	14-283	189
	T4	55-56	UCIAMS64583	Sph stems/Picea leaf frs	325±20	319-488	410
	T4	70-71	UCIAMS57420	Sph stems	1250±15	1095-1293	1211
MOS	C	40-41	UCIAMS57424	Sph stems	355±15	325-522	444
	C	70-71	UCIAMS54958	Sph stems	1270±25	1095-1313	1223
	C	95-96	UCIAMS64586	Sph stems	1990±20	1837-1985	1924
	C	108-109	UCIAMS67515	Sph stems	2065±25	1976-2119	2043
	C	120-121	UCIAMS54959	Sph stems	2225±25	2157-2335	2237
	C	136-137	UCIAMS64588	Sph stems	2490±20	2478-2728	2591
	C	172-173	UCIAMS54960	Sph stems	3275±25	3405-3604	3506
	C	224-225	UCIAMS54961	Sph stems	4185±25	4609-4863	4739
	C	246-247	UCIAMS57426	Sph stems	4740±15	5333-5634	5534
	C	296-297	Beta223744	Sph stems	6200±40	6936-7237	7072
	L1	55-56	UCIAMS65378	Charcoal frs	190±15	149-310	229
	L1	77-78	UCIAMS65385	Charcoal frs	1260±20	1100-1282	1216
	L1	89-90	UCIAMS65389	Charcoal frs	1840±20	1705-1907	1781
	L1	117-119	UCIAMS65386	Charcoal, Cyp seeds	3625±20	3067-3983	3897
	L1	141-144	UCIAMS65375	Charcoal, Cyp seeds; Picea leaf frs	3070±20	3418-4734	Rejected
	L1	167-170	UCIAMS43474	Eric leaf frs; Cyp seeds	6420±20	6799-7749	7340
	L2	160-164	UCIAMS43475	Picea leaf frs	5655±20	6401-6484	6443
	L3	211-213	UCIAMS40364	Sph stems	5755±20	6491-6630	6561
	L4	51-52	UCIAMS57425	Sph stems	455±15	502-550	545
	L4	73-74	UCIAMS58642	Sph stems	2165±20	2097-2285	2130
	L4	97-99	UCIAMS58641	Eric/Larix leaf frs	2750±20	2790-2911	2852
	L4	136-137	UCIAMS58640	Eric/Larix/Picea leaf frs	3835±15	4162-4326	4322
	L4	169-170	UCIAMS43476	Sph stems	4670±20	5318-5457	5323
	T4	24-25	UCIAMS64584	Charcoal	165±20	170-476	210
	T4	67-68	UCIAMS64589	Charcoal	2670±20	2342-2834	2794
	T4	83-84	UCIAMS65379	Charcoal	4055±20	3997-4603	4407
STE	C	45-46	UCIAMS54962	Sph stems	105±30	75-260	116
	C	67-68	UCIAMS58645	Picea leaf frs	600±20	548-646	620
	C	79-80	UCIAMS64589	Sph stems	1175±20	1039-1166	1112
	C	98-99	UCIAMS54963	Sph stems	1715±25	1568-1697	1578

C	124-125	UCIAMS65381	Sph stems	2445±20	2351-2663	2461
C	160-161	UCIAMS54964	Sph stems	3255±30	3316-3521	3505
C	179-180	UCIAMS67514	Sph stems	3415±25	3588-3671	3597
C	194-195	UCIAMS58644	Sph stems	3125±15	3654-3763	Rejected
C	201-202	UCIAMS65382	Sph stems	3485±20	3757-3912	3805
C	223-224	UCIAMS54965	Sph stems	3960±30	4283-4431	4362
C	239-240	UCIAMS58643	Sph stems/Picea leaf frs	3975±15	4422-4685	4424
C	285-286	UCIAMS40360	Sph stems	6225±20	6731-7219	7007
L1	161-165	UCIAMS40361	Sph stems; Cyp seeds; Picea leaf frs	6215±20	7016-7238	7127
L2	44-45	UCIAMS67506	Sph stems	265±25	269-436	316
L2	68-69	UCIAMS67507	Sph stems	1195±25	1031-1232	1123
L2	98-99	UCIAMS67508	Sph stems	2380±25	2329-2588	2399
L2	142-143	UCIAMS67509	Sph stems	3200±25	3360-3471	3415
L2	180-181	UCIAMS67510	Sph stems	3820±25	4108-4363	4207
L2	212-213	UCIAMS67511	Sph stems	4465±25	4977-5288	5182
L2	244-246	UCIAMS40362	Sph stems	5760±20	6412-6725	6550
L3	135-139	UCIAMS43477	Picea/Eric leaf frs	5215±20	5924-5995	5960
L4	35-36	UCIAMS65384	Sph stems	135±20	70-287	224
L4	48-49	UCIAMS67512	Charcoal frs	945±25	778-967	867
L4	63-64	UCIAMS65380	Charcoal; Picea leaf frs	2455±20	2343-2708	2524
L4	84-85	UCIAMS67513	Charcoal frs	3540±25	3703-3948	3826
L4	125-126	UCIAMS65376	Charcoal, Picea/Eric leaf frs	5490±20	6189-6339	6288
L4	174-176	UCIAMS40363	Sph stems; Larix leaf frs; Cyp seeds	6185±20	6976-7345	7090
T4	46-47	UCIAMS65383	Sph stems	165±20	135-328	225
T4	63-64	UCIAMS65377	Charcoal frs	1660±20	1442-1789	1557

Table 1.1. Radiocarbon datings for dated samples, listed per peatland and core. Cores were separated according to their position: C = central core; L = lateral core; T = transitional core. Dated vegetation types were abbreviated: Sph = *Sphagnum* spp.; Eric = Ericaceae; Cyp = Cyperaceae.

	LLC	MOS	STE
Surface area (km^2)	2.241	2.672	1.722
Mean peat thickness (m)	2.50	1.82	1.87
Peat volume (10^6 m^3)	5.614	4.864	3.225
Mean C density (kg m^{-3})	42.5 (0.363)	44.5 (0.427)	45.8 (0.559)
250 cal BP - present	32.2 (0.600)	33.5 (0.705)	32.8 (0.993)
mineral base - 250 cal BP	46.0 (0.387)	48.9 (0.438)	51.6 (0.556)
Mean C mass per area (kg m^{-2})	106.4	81.0	85.8
Total C mass (10^6 kg)	242.4	216.8	148.5
Mean basal age (cal BP)	5664	5576	5636
C accumulation rate (g m^{-2} yr^{-1})			
mineral base - present	18.9	14.4	15.2
250 cal BP - present	53.9	55.8	59.4

Table 1.2. Peatland characteristics. Standard errors are in parentheses.

CHAPTER II

HOLOCENE CARBON ACCUMULATION RATES FROM THREE OMBROTROPHIC PEATLANDS IN BOREAL QUEBEC, CANADA: IMPACT OF CLIMATE-DRIVEN ECOHYDROLOGICAL CHANGE

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Article in press in The Holocene.

Résumé

La compréhension des processus qui influencent l'accumulation du C dans les tourbières est essentielle afin d'anticiper les changements potentiels dans le cycle global du C en lien avec les changements climatiques. L'identification des facteurs qui influencent l'accumulation du C est relativement simple à l'échelle annuelle, cependant, à l'échelle séculaire ou millénaire une interaction complexe entre le climat, l'hydrologie, les processus phytoécologiques internes et les perturbations complique l'identification des facteurs déterminants. Afin de mieux comprendre l'impact de ceux-ci, les taux d'accumulation de C durant l'Holocène ont été quantifiés dans trois tourbières ombrotrophes au Québec boréal, au nord-est du Canada (52°N, 75-76°O).

Les analyses de densité sèche et de la perte-au-feu de la tourbe, en combinaison avec les chronologies au radiocarbone (^{14}C) ont permis la reconstruction des taux d'accumulation du C durant l'Holocène. Des reconstructions des variations d'assemblages végétaux et de la nappe phréatique ont été obtenues à partir de l'analyse des macrorestes végétaux et des thécamoebiens.

L'accumulation de la tourbe a commencé vers 7520 cal BP dans la région d'Eastmain, avec des taux moyens variant entre 14,9 et 22,6 g m^{-2} an^{-1}. Les taux d'accumulation élevés au cours de l'Holocène moyen sont associés à la présence des sphaignes de la section *Acutifolia*, tandis que les périodes de ralentissement de l'accumulation de l'Holocène récent sont caractérisées par une végétation ligneuse importante et de cypéracées. Les résultats montrent que l'accumulation du C a varié de façon importante entre les sites, indiquant que les contextes géomorphologiques, de variations hydrologiques, ou encore des évènements ponctuels tel que les feux, ont pu atténuer l'influence générale du climat. Les reconstructions de nappe phréatique présentent plusieurs épisodes de bas niveau de nappe phréatique depuis 3000 cal BP, suggérant que des conditions froides et sèches ont pu influencer une diminution des taux d'accumulation du C.

Étant donné l'intensité de ces fluctuations de la nappe phréatique, nous proposons l'hypothèse que l'occurrence de périodes de gel de l'ordre de quelques années ou décennies pourraient avoir contribué à la sécheresse apparente de la surface de la tourbière. Les projections climatiques indiquant pour la région des conditions généralement plus chaudes et humides, favorisent une plus grande stabilité des sphaignes de la section *Acutifolia*, et donc une augmentation du potentiel de la séquestration du C.

Abstract

Understanding the processes controlling peatland carbon (C) sequestration is critical to anticipate potential changes in the global C cycle in response to climate change. Although identification of these factors may be relatively straightforward on seasonal timescales, at centennial to millennial timescales complexities arise due to interactions between climate, vegetation, hydrology and long-term ecological processes. To better understand the factors controlling long-term C accumulation, Holocene rates of C sequestration were quantified from three pristine ombrotrophic peatlands in boreal Quebec, northeastern Canada (52°N, 75-76°W).

Bulk density and loss-on-ignition analyses, combined with radiocarbon dating and age-depth modelling, were used to estimate long-term apparent rates of carbon accumulation. Past changes in vegetation and water table depth were obtained from plant macrofossil and testate amoeba analysis.

Earliest regional peat accumulation started ~7520 cal BP, with long-term rates of C sequestration varying between 14.9 and 22.6 g m^{-2} yr^{-1}. High C sequestration rates occurred during the mid-Holocene when relatively stable *Sphagnum* section *Acutifolia* communities were present, while low rates were found during the cooler late-Holocene when Cyperaceae and ligneous vegetation were more dominant. However, C sequestration was highly variable among cores, implying that local topography, geomorphology and hydrology, or disturbance factors such as fire, mediate the influence of climate on C accumulation. Reconstructed water-table depths reveal several dry shifts since 3000 cal BP, suggesting that episodic cold and dry conditions during the late-Holocene may have contributed to lower C sequestration rates.

Given the intensity of the water table shifts at these times, we hypothesize that recurrent episodes of frozen subsurface peat might have intensified surface drying. As projected by climate scenarios, anticipated warmer and wetter conditions may lead to greater stability of hummock *Sphagna* cover and increased C sequestration potential in boreal peatlands.

2.1. Introduction

Peatlands are environments where anaerobic conditions restrict decomposition, leading to the accumulation of organic matter through time. Over millennia, a small net imbalance between net primary production (NPP) and decomposition of northern peatland biomass has resulted in a global stock of organic carbon (C) attaining ~547 Pg (Yu *et al.*, 2010). Peatland C dynamics have affected climate change during the past millennia, by providing long-term positive and negative feedbacks on atmospheric greenhouse gas concentrations (Frolking and Roulet, 2007).

Besides important long-term C sinks, ombrotrophic peatlands constitute high-resolution archives of climatic change due to their sensitivity to variations in atmospheric moisture balance (Charman, 2002). Consequently, stratigraphic surveys of both past C sequestration and reconstructions of past hydroclimate from these systems may provide insight into future projections of C dynamics in response to climatic change. In this study, we quantified long-term C accumulation rates and investigated past variations in hydroclimatic conditions and plant assemblages at three peatlands in boreal Quebec, northeastern Canada.

Factors regulating the strength of the peatland C sink vary at different timescales (Yu, 2006b). Short-term relative rates of NPP and decomposition, quantified by net ecosystem productivity (NEP), are primarily linked to shifts in water table depth (WTD) and peat surface temperature (e.g. Bubier *et al.*, 2003 ; Pelletier, Garneau and Moore, in review ; Silvola *et al.*, 1996). However, on millennial timescales, changes in peatland vegetation and WTD, and thus C sequestration, are driven by a combination of allogenic and autogenic factors. Globally, climate regime may be the principal allogenic factor having a concomitant effect on both vegetation and hydrology (Barber *et al.*, 1994), although fire, permafrost or geological processes (e.g. isostatic uplift) may be as important in certain regions (Glaser *et al.*, 2004 ; Pitkänen, Turunen and Tolonen, 1999 ; Turetsky *et al.*, 2007). Autogenic factors affecting C sequestration may include surface microtopography dynamics (Eppinga *et al.*, 2009 ; Swanson, 2007) and hydrology-induced limits to vertical peat growth (Belyea and Clymo, 2001 ; Yu *et al.*, 2003).

Holocene variations in C sequestration have been reconstructed in various peatland types and climatic regimes. Most studies in northern peatlands show a slowdown of net

vertical C accumulation towards the late-Holocene (e.g. Turunen and Turunen, 2003 ; Turunen *et al.*, 2001 ; Yu, 2006a), although one would expect an apparent net acceleration, as the duration of decay processes is higher in older peat (Clymo, 1984). Late-Holocene climatic cooling, affecting peatland ecohydrology and probably NPP, may have been responsible for declining rates of C sequestration. Nevertheless, long-term surface drying driven by autogenic factors associated with peatland development may have also contributed to late Holocene declines in C sequestration.

In this study, we quantified variations in Holocene C accumulation from three pristine boreal peatlands located in the Eastmain region, northeastern Canada. In addition, by directly quantifying past peatland vegetation and water table fluctuations, linkages among C accumulation, dominant vegetation and hydrological conditions were investigated. Finally, we also used our data to infer patterns of regional climate change based on replicate variations within cores.

2.2. Study region

Peat cores were recovered from the deepest parts of the pristine ombrotrophic peatlands Lac Le Caron (LLC; 52°17'15"N / 75°50'21"W), Mosaik (MOS; 51°58'55"N / 75°24'06"W) and Sterne (STE; 52°02'37"N / 75°10'23"W), located in the Eastmain river watershed in the boreal region of the James Bay lowlands, Quebec, Canada (Fig. 2.1). Regional mean annual temperature is -2.1 ± 0.2°C (January: -22.0 ± 0.5°C; July: 14.6 ± 0.2°C), mean precipitation is 735 ± 12 mm, of which about one third falls as snow, and the mean value for growing degree days above 0°C is 1763 (interpolated means and standard errors of 1971-2003 NLWIS data; Hutchinson *et al.*, 2009). Further details on the study region and peatlands can be found in van Bellen *et al.* (2011).

2.3. Material and methods

2.3.1 The use of multiple proxies

Paleoenvironmental reconstructions are preferentially based on multiple proxies, as independent records may prevent spurious reconstructions from a single proxy (Blaauw, Bennett and Christen, 2010 ; Charman, Hendon and Packman, 1999). Multiple proxies are particularly valuable in retrospective studies of peatland C dynamics, because rates of C accumulation are a product of multiple interacting factors (e.g., climate, vegetation, hydrology). Shifts in peat-forming vegetation are often related to changes in C sequestration, and plant macrofossils allow the identification of these past plant communities. In addition, plant macrofossils constitute an archive of moisture balance conditions, as changes in vegetation are primarily driven by hydroclimatic shifts if disturbance is absent. Likewise, assemblages of testate amoebae are recognized as good indicators of water table depth (WTD) in *Sphagnum*-dominated peatlands, especially when transfer functions are applied (e.g. Booth, 2008 ; Charman, Blundell and ACCROTELM members, 2007 ; Mitchell, Charman and Warner, 2008). Testate amoebae are highly sensitive to short-term (seasonal) and long-term (decadal) variations in WTD (Booth, 2010 ; Warner, Asada and Quinn, 2007).

2.3.2. Fieldwork

One core was recovered from the deepest part of each peatland (van Bellen *et al.*, 2011), using a Box corer (10×10 cm width) to sample the upper 1 m and Russian peat samplers (4.5- to 7.5-cm diameter) for deeper peat. Sampled monoliths were wrapped in plastic, transferred to polyvinyl chloride tubes and stored at 4°C until analysis.

2.3.3. Laboratory work

Prior to specific treatment, cores were sliced into contiguous 1 cm subsamples. Long-term apparent carbon accumulation rates (LORCA) were calculated from bulk density and loss-on-ignition (LOI) analyses. The amount of organic matter (OM) was defined as the product of bulk density and LOI (Dean, 1974). Bulk density was determined from contiguous 1 cm^3 subsamples after drying for 16 hours at 105°C. Subsequently, LOI analysis was performed at 550 C for 3.5 hours (Heiri, Lotter and Lemcke, 2001). The amount of organic C was calculated assuming a mean of 0.5 g C g^{-1} OM (Turunen *et al.*, 2002).

Plant macrofossil analysis was performed at 4 cm resolution. Subsamples (4 cm^3) were gently heated in distilled water with addition of 5% KOH and carefully rinsed through a 125-µm sieve. A binocular microscope (×10-×40) was used to identify peat-forming vegetation. Vegetation types were estimated as volume percentages of the 4-cm^3 subsample; seeds, charcoal fragments and *Cenococcum* sclerotia were counted and acarid mite/insect fragments were scaled 1 to 10 (1=rare, 10=abundant). *Sphagnum* sections were determined using a microscope (×40-×100) based on branch leaf characteristics because stem leaves, which are often essential for the identification to species level, were rarely found. A plant macrofossil reference collection (Garneau, 1995) was used for identification; vegetation terminology is from Marie-Victorin (1995) and *Sphagnum* taxonomy follows Ayotte and Rochefort (2006).

Testate amoeba assemblages were analyzed at the same resolution as plant macrofossils. Subsamples (1 cm^3) were gently boiled in distilled water according to the protocol described by Hendon and Charman (1997) and *Lycopodium* (or to some samples *Eucalyptus*) was added as an exotic marker to permit the calculation of test concentrations. The material between 355 and 15 µm was retained after sieving. A total of 150 specimens was counted in most samples, although for a few samples this count total could not be achieved. Samples containing less than 75 tests were considered unreliable, which is broadly consistent with the recommendation of Payne and Mitchell (2008) that counts of 50 may be adequate for some sites and counts of 100 are likely to be sufficient for most sites. Abundance of species was expressed as a percentage of the total tests counted. The common rotifer *Habrotrocha angusticollis* was included in the percentage calculations. Taxonomy

follows Charman, Hendon and Woodland (2000) except as modified by Booth (2008). Plant macrofossil and testate amoeba diagrams were created using C2 and plotted against depth (Juggins, 2007). Zones were identified on a visual basis, according to patterns in dominant vegetation, testate amoebae and C accumulation.

A total of 34 samples were submitted to Keck-CCAMS facility (Irvine, USA) and Beta Analytic Inc. (Miami, USA) for dating by ^{14}C accelerator mass spectrometry (AMS). Sample material was preferentially *Sphagnum* stems (Nilsson *et al.*, 2001), although in some cases *Picea* or Ericaceae leaf fragments were dated (Table 2.1).

2.3.4. Data analysis

Radiocarbon ages were calibrated using the IntCal04 calibration curve (Reimer *et al.*, 2004) within the Bchron software package in R (Haslett and Parnell, 2008 ; http://www.r-project.org; R Development Core Team, 2009). One radiocarbon date on the STE core was identified as an outlier and was therefore omitted when developing the age-depth model. Bchron was further used for age-depth modelling, considering vertical accumulation as a continuous monotonic process applying piecewise linear interpolation. This approach includes creation of the 95% 'highest density region' of any level in the sediment record, thereby allowing estimations of the age uncertainty of any horizon within the cores. All ages were expressed as calendar years before present (BP = before AD 1950). Age-depth models were based on these calibrated ages and the age of the peatland surface, established at -56 cal BP (i.e. AD 2006). Age estimates were rounded to the nearest decade.

A transfer function based on 650 modern assemblages from peatlands in eastern and mid-continental USA was used to infer quantitative WTD from fossil testate amoeba assemblages (Booth, 2008). The widespread distribution of many testate amoeba taxa, and the similarity in the ecology of taxa among regions, suggests that the calibration dataset of Booth (2008) should provide reliable water-table depth estimates in boreal Quebec, particularly along a relative water-table depth gradient (e.g. Booth, 2008 ; Booth, Sullivan and Sousa, 2008 ; Charman and Warner, 1997). A weighted average-partial least squares model was used, as it showed optimal r^2 and RMSEP values although application of other

models resulted in similar reconstructions (Booth 2008). Water-table depth reconstructions were expressed as raw values and were also detrended by a LOWESS filter with a 50% smoothing factor. Smoothing was performed to remove long-term trends caused by autogenic factors (i.e. lateral expansion and consequential drainage; Charman *et al.*, 2006) to highlight hydrological changes most likely linked to hydroclimate variability. All records were standardized by calculating the difference from the mean of all observations within each core, divided by its standard deviation (i.e. z-scores). Important events of extremely high or low water table levels were visually linked, taking account of the uncertainty associated with each chronology.

Apparent C accumulation rates were defined as the product of sample C density and vertical accumulation rate (Tolonen and Turunen, 1996) and plotted as a function of age to facilitate inter-site comparisons. However, short-term shifts in C accumulation rates should be interpreted with caution, as our age-depth models were not highly resolved and thus generally failed to detect decadal to centennial shifts in vertical accumulation rates. Generally, episodes of slow vertical accumulation coincided with the presence of denser peat. As a result, decadal to centennial episodes of slow vertical accumulation not detected by age-depth models may be displayed as unrealistic peaks in C accumulation rate (i.e. post-fire; Kuhry, 1994). To account for these flaws in the age-depth models, a LOWESS smoothing was applied to the C accumulation data with the span set as the inverse of the number of radiocarbon dates per core.

2.4. Results

2.4.1. LLC bog ecohydrological reconstructions

Peat accumulation started around 7520 cal BP under minerotrophic conditions followed by an early shift to ombrotrophy around 7210 cal BP as indicated by the presence of *Sphagnum* section *Acutifolia* (Fig. 2.2, Table 2.2). Stratigraphic analyses show an important stable presence of *Sphagna* and high concentrations of testate amoebae throughout the deeper parts of the profile (zones LLC2, LLC3 and LLC4; Fig. 2.2). From 4170 cal BP vegetation

assemblages alternate between *Sphagnum*, Cyperaceae and Ericaceae dominance (zone LLC5). This trend is accentuated between 2960 and 330 cal BP, when repeated high-magnitude fluctuations in WTD (2-30 cm below surface) are registered through shifting testate amoeba assemblages of *Difflugia pulex*, *Amphitrema flavum*, *Amphitrema wrightianum*, *H. angusticollis*, *Heleopera sphagni* and *Hyalosphenia subflava* (zones LLC6 and LLC7; Fig. 2.2 and Table 2.2). In addition, large charcoal fragments (>2 mm) indicate repeated fires on the coring site or in its vicinity, concentrated between 2520 and 390 cal BP. Well-preserved *Sphagnum* section *Acutifolia* peat accumulated since 330 cal BP, while a shift to more humid conditions is apparent in the uppermost section of the stratigraphy (zone LLC8).

2.4.2. LLC bog carbon accumulation and biohydrological proxies

The LLC age-depth model shows a convex shape, indicating that vertical peat accumulation slowed down during the late-Holocene, except for the uppermost part that is partially represented by the acrotelm (Fig. 2.3a). The reconstruction of C accumulation in LLC shows that rates have varied considerably during the Holocene (Fig. 2.4) with a mean of 22.6 g m^{-2} yr^{-1}. Despite high decomposition, C accumulation rates were high during the fen phase (zone LLC1; Table 2.2). Besides this period, rapid C accumulation occurred at the upper part of zone LLC2 (~5700-5040 cal BP), characterized by moderate water tables and *Sphagnum* section *Acutifolia*. C accumulation slowed down during zones LLC4 and LLC5 (4420-2960 cal BP), attaining minima of 9-11 g m^{-2} yr^{-1} in zone LLC6 (2960-1250 cal BP). This period of limited C sequestration was characterized by high mean water tables interspersed with episodic dry surfaces (Fig. 2.4). The corresponding tendencies of dry-indicating macrofossils and the inferred WTD show that extreme fluctuations in surface hydrology have occurred during the period of slow accumulation, although low temporal sample resolution between 4000 and 1000 cal BP somewhat complicates the interpretation. A large increase in C sequestration rates started at the onset of zone LLC7 at 1250 cal BP. The acceleration in C accumulation may well be related to the establishment of *Sphagnum* section *Acutifolia*, mediated by a climatic warming enhancing primary production. During the last

millennium, C sequestration rates varied only slightly between 25 and 30 g m^{-2} yr^{-1} and, except for diminishing *Sphagnum* presence around 400 cal BP, vegetation remained relatively stable.

2.4.3. MOS bog ecohydrological reconstructions

Plant macrofossil analyses suggest that *Sphagna* were less important in MOS than LLC (Fig. 2.2 and 2.5). *Sphagnum* section *Acutifolia*, implying relatively dry local conditions, has been present during the Holocene, yet collapsed frequently at the expense of other *Sphagnum* sections, Cyperaceae or Ericaceae (Fig. 2.5, Table 2.3). Following peat inception at 7070 cal BP, fen conditions persisted until 5450 cal BP with water tables at the surface as indicated by high amounts of *A. wrightianum*, *D. pulex* and *A. flavum* (zone MOS1). Nevertheless, WTD reconstructions should be interpreted with caution as less than 75 specimens were counted in some samples. The mid-Holocene (zones MOS2 and MOS3) is characterized by *Sphagnum*-Cyperaceae and *Sphagnum*-Ericaceae peat with *D. pulex* and *A. flavum*. Zone MOS4, starting around 1940 cal BP, shows a general increase in Ericaceae, less dominant *Sphagnum* presence and sporadic *Pleurozium schreberi* around 1890 and 1000 cal BP. Episodic low water tables are encountered during this period, identified by *H. subflava*. Testate amoeba assemblages at these times are unusual, with large amounts of species associated to both rather wet (*A. discoides* and *A. flavum*) and dry optima (*D. pulex*, and *H. subflava*). Finally, a steady increase in *Sphagnum* section *Cuspidata* started ~240 cal BP continuing until present-day, although a dry shift was marked by *H. subflava* presence around 150 cal BP (zone MOS5, Table 2.3).

2.4.4 MOS bog carbon accumulation and biohydrological proxies

The MOS age-depth relationship is close to linear, lacking the slowdown of vertical accumulation that was evident at LLC (Fig. 2.3b). Nevertheless, mean apparent C accumulation rate for MOS is substantially lower at 15.4 g m^{-2} yr^{-1}. Although MOS shows

long-term variations in accumulation rates throughout the Holocene, trends are less clear than in LLC (Fig. 6). In accordance, the vegetation record shows no clear long-term tendencies. C accumulation rates remained relatively stable between 10 and 20 g m^{-2} yr^{-1} from the start of local peat accumulation until the first half of zone MOS3 (~2700 cal BP), followed by highest accumulation rates (17-32 g m^{-2} yr^{-1}) until the start of zone MOS4 at 1940 cal BP. This peak is associated with an important presence of *Sphagnum* section *Acutifolia* and WTD averaging 9-14 cm, while decomposition, indicated by the amounts of unidentifiable organic matter (UOM), was low. The lowest C accumulation rates of 12 g m^{-2} yr^{-1} and considerably fluctuating water table levels are reconstructed for the period 1920-1180 cal BP.

2.4.5. STE bog ecohydrological reconstructions

The entire sequence of STE has high amounts of Ericaceae/wood peat and, except for the upper 60 cm, low presence of *Sphagna* (Fig. 2.7). As in LLC and MOS, *A. flavum*, *D. pulex* and *H. subflava* are the most frequent testate amoebae (Fig. 2.7). In accordance with the dominance of ligneous vegetation, reconstructed water-table depths were generally lower than LLC and MOS and hydrological fluctuations showed high amplitudes. Peat accumulation started from 7010 cal BP under minerotrophic conditions. High decomposition has resulted in very low countable quantities of testate amoebae for zone STE1 (Table 2.4). Of the three peatlands, STE shows the latest transition to ombrotrophy around 4410 cal BP. *Sphagnum* section *Acutifolia* established during ombrotrophication and remained relatively important throughout zone STE2 until 3230 cal BP. At the onset of STE3, *Sphagnum* section *Acutifolia* decreased, while Cyperaceae and *H. subflava* became more abundant. Extremely dry surface conditions, starting 1090 cal BP, distinguish zone STE4 from adjacent zones, while water tables remained between 25 and 37 cm below the surface from 990 to 450 cal BP. During this period, *H. subflava* and *D. pulex* dominate the assemblages, and the vegetation cover was characterized by Ericaceae and *Picea*. A consecutive hydrological shift centered around 220 cal BP allowed an accumulation of well-preserved *Sphagnum* peat (zone STE5).

2.4.6. STE bog carbon accumulation and biohydrological proxies

Peat accumulation in STE shows highly variable rates, with rapid accumulation during the mid-Holocene (Fig. 2.3c). With the exception of high rates during the mid-Holocene, C has been sequestered continuously at 10-20 g m^{-2} yr^{-1}, resulting in a Holocene LORCA of 14.9 g m^{-2} yr^{-1} (Fig. 2.8). The fen phase is associated with low sequestration rates (zone STE1). The subsequent mid-Holocene period of enhanced C accumulation shows abundance of *Sphagnum* section *Acutifolia* (Fig. 2.8; zone STE2). The two peaks observed around 4410 and 3650 cal BP are principally the result of reconstructed phases of rapid vertical accumulation of peat formed under dry conditions. Improbable sharp increases in C accumulation rates, associated with this highly decomposed peat, is likely the result of flaws in the age-depth model. Nevertheless, C sequestration rates were relatively constant during the major part of the Holocene. The 1110-620 cal BP period of slow accumulation coincides with declining water table levels and a persistent extremely dry peatland surface (WTD attaining 37 cm; zone STE4), while *Sphagnum* presence generally diminished and ligneous vegetation increased.

2.4.7. Regional paleohydrological variability: climatic influence

The compiled paleohydrological reconstructions from the ombrotrophic portion of the three sites cover the last 4300 years. Although the uncertainty of our age-depth models complicates precise correlations among the sites, numerous hydrological shifts are observed since 3000 cal BP at all sites. Within the uncertainty of our chronologies, at least seven dry and four wet events were probably registered regionally (Fig. 2.9). Dry conditions were centered around 2790, 2500, 1910, 1290, 770, 400 and 140 cal BP, with the dry events at 770 and 400 cal BP particularly well recorded by the three sites. Regional wet shifts appear to have occurred around 2330, 1390 and 930 cal BP, and all sites indicate wetter conditions after ~150 cal BP (Fig. 9). In addition, MOS and STE records show increasing *Sphagnum* section *Cuspidata* during the last century, strengthening the interpretation of the testate amoeba record.

2.5. Discussion

2.5.1. Linkages among long-term C sequestration, vegetation and hydrology

Although we may assume a uniform climatic history for the Eastmain region, the apparent rates and the timing of shifts in C accumulation are highly variable between peatlands. Mean Holocene LORCA has been higher in LLC with 22.6 g m^{-2} yr^{-1} compared to MOS and STE, with 15.4 and 14.9 g m^{-2} yr^{-1} respectively. To be able to estimate causes for differential C sequestration patterns, the entire accumulation history and environmental context of each core should be taken into account. C accumulation rates have been higher during the early bog phases of LLC (zone LLC2, 7210-5040 cal BP), when a combination of a persistent *Sphagnum* section *Acutifolia* cover and probably favourable mid-Holocene climatic conditions (Kaufman *et al.*, 2004 ; Viau *et al.*, 2006) allowed a rapid vertical accumulation. High rates of accumulation may have resulted of a limited decomposition due to the decay-resistant nature of *Sphagnum* section *Acutifolia*. Although climatic conditions would have been equally favourable for MOS and STE, *Sphagnum* could not yet become dominant in the local vegetation during this period as minerotrophic conditions persisted until 5450 cal BP in MOS and 4410 cal BP in STE. Globally, fens may be less efficient C sinks than bogs (Turunen *et al.*, 2002), although C accumulation under minerotrophic conditions was high in LLC (>30 g m^{-2} yr^{-1}) and low in STE (7-14 g m^{-2} yr^{-1}). The lack of *Sphagna* during minerotrophic conditions likely contributed to the slow C sequestration in MOS and STE.

Of all the peatlands in this study, the most important temporal variations in C sequestration are observed in LLC (Fig. 2.4), with a mid-Holocene (5790-4170 cal BP) period of rapid accumulation and a late-Holocene (4170-1250 cal BP) period of slow accumulation. The vegetation record shows a similar subdivision with a relatively stable *Sphagnum* cover that collapsed repeatedly between 4000-1300 cal BP. The compilation of all stratigraphic data shows that C accumulation rates declined simultaneously with a rise in mean water table of ~10 cm at the onset of LLC3 (5040 cal BP), although vegetation composition remained stable until 4170 cal BP (Fig. 2.6). This may indicate that diminishing C sequestration was driven primarily by a shift in WTD, that was exacerbated when *Sphagna*

finally collapsed after 4170 cal BP. An initial persistence of relatively high water tables may have exerted a stress on *Sphagnum* section *Acutifolia*, followed by a replacement by other *Sphagnum* sections and Cyperaceae and a continued slowdown of C accumulation.

Furthermore, slowest C sequestration in LLC between 2960 and 1250 cal BP was characterized by fluctuating water tables (zone LLC6). Extremely low water tables cause a slow passage of organic matter through the acrotelm, resulting in a high decay potential. Moreover, hummock *Sphagna* NPP may decrease due to shading as Ericaceae cover increases (Heijmans *et al.*, 2001). On the other hand, extremely high water tables may enhance microtopographical gradients by expanding wet hollows and increasing hummock height (Swanson, 2007). In addition, wet bog surfaces are associated with bryophytes that are sensitive to decomposition. Fluctuating water tables may have caused testate amoeba assemblages consisting of *H. subflava* and *A. discoides* as observed in MOS4. *A. discoides* is typically most abundant in wet habitats whereas *H. subflava* is characteristic of drier habitats (Booth, 2008), but the co-dominance of the two has been observed in several studies and may be the result of high seasonal-to-interannual variability in moisture (Booth, Sullivan and Sousa, 2008 ; Charman, Blundell and ACCROTELM members, 2007 ; Sullivan and Booth, in review). Indeed, maximum C accumulation has been associated with an acrotelm thickness of 10-15 cm, typical for lawns or low hummocks (Belyea and Clymo, 2001).

Different vegetation and fluctuating water tables may partially explain the differences in C sequestration during the bog stages within LLC as well as differences between the sites. Previous research showed that LLC bog has a strong surface topography gradient at ~5 m km^{-1} that may have supported efficient general drainage (van Bellen *et al.*, 2011), which may have contributed to more stable ombrotrophic conditions and hummock *Sphagna* cover.

Neoglacial cold conditions could have further limited accumulation in LLC by lowering primary production and limiting the duration of the growing season (Mauquoy *et al.*, 2002a). Cooling conditions leading to decreased NPP have been reported from tree stands close to the Eastmain region after 4650 cal BP (Arseneault and Sirois, 2004). Although decreasing C accumulation rates are perceptible between 1430 and 1180 cal BP in MOS and 1070 and 660 cal BP in STE, climate-driven depressed primary production may have been less important in these peatlands, as C sequestration was already limited due to the absence of an important *Sphagnum* cover.

The long-term effect of climate on NEP may be illustrated by large spatial patterns of peatland C storage (Yu, Beilman and Jones, 2009). C accumulation rates have an optimum in the 'temperature-precipitation space' at 0-2.5°C mean annual temperature (Yu, Beilman and Jones, 2009) and generally diminish following nordic and oceanic gradients in Quebec, i.e. from a cool and moist to a cold and dry regime. Thus, it is likely that Neoglacial conditions have limited C sequestration in peatlands within the northern Quebec climatic context. Studies on boreal and subarctic Quebec peatland C accumulation indeed show a general slowdown of accumulation during the Neoglacial cooling. Beaulieu-Audy *et al.* (2009) showed declining rates of peat accumulation starting around 4000 cal BP in northern Quebec. Providing late-Holocene records, Loisel and Garneau (2010) obtained minimum C accumulation of <10 g m^{-2} yr^{-1} from two cores in MOS peatland during the 2500-1000 cal BP period.

Corresponding dry and wet shifts registered in Eastmain region peatlands may imply a climatic forcing that limited C sequestration. However, inconsistencies between the studied sites show that internal factors are likely to have mediated external forcing of C sequestration at varying intensities. Reconstructions of Eastmain peatland C accumulation at the ecosystem scale have already shown that long-term peatland development is affected by site-specific factors such as basin topography (van Bellen *et al.*, 2011).

2.5.2. Late-Holocene climate change in the Eastmain region

The replicated shifts in peatland WTD may indicate an enhanced climatic influence on ecosystem functioning since 3000 cal BP. Ombrotrophic peatland hydrology is directly linked to the balance between precipitation and evapotranspiration, although the respective importance of precipitation and temperature to this balance may vary between regions and climatic regimes and even in the course of peatland development (Booth, 2010 ; Charman *et al.*, 2009). The combined records of water table fluctuations (Fig. 9) show that during the late-Holocene (3000 cal BP-present) hydroclimatic variations were important, possibly indicating less stable conditions than during the mid-Holocene (7000-4000 cal BP). However, comparison of the two periods is complicated for several reasons. First, we do not have a

combined record that covers the entire mid-Holocene. In addition, different stages in long-term peatland development may have differential sensitivities to external factors. Finally, the comparison of the frequency, duration and timing of hydrological shifts between mid- and late-Holocene is complicated by lower temporal resolution of the late-Holocene record and the larger age uncertainty for mid-Holocene shifts. Relatively low-resolution analyses, with one sample each 4 cm, may imply that some dry shifts may have been missed. In an eastern Newfoundland peatland, Hughes *et al.* (2006) found high-frequency hydrological variability during the mid-Holocene, possibly related to global climatic instability. However, regional differences in hydroclimatic context and concomitant differential sensitivity to climatic driving factors likely hamper direct comparisons of Newfoundland and James Bay region records.

In northeastern Canada, cooling climatic conditions are associated with the intrusion of dry Arctic air masses (Carcaillet and Richard, 2000 ; Girardin *et al.*, 2004). Hence, past cooling events may have been concomitant with a decrease in both annual precipitation and potential evaporation, resulting in contrasting effects on peatland water tables. Eastern Canadian climate history can be summarized by a mid-Holocene thermal maximum (Kaufman *et al.*, 2004 ; Viau *et al.*, 2006) and a gradual cooling trend that started between 5000 and 4000 cal BP (Arseneault and Sirois, 2004 ; Filion, 1984 ; Kerwin *et al.*, 2004). This Neoglacial period affected the distribution of boreal ecozones with more severe fire events (Ali *et al.*, 2008 ; Ali, Carcaillet and Bergeron, 2009 ; Arseneault and Sirois, 2004 ; Asselin and Payette, 2005 ; Payette and Gagnon, 1985), whereas permafrost gradually developed in subarctic Quebec peatlands (Bhiry, Payette and Robert, 2007 ; Bhiry and Robert, 2006 ; Lamarre, 2010). This period may have been characterized by humid climatic conditions as shown by regional increases in lake levels (Miousse, Bhiry and Lavoie, 2003 ; Payette and Filion, 1993). In addition to Neoglacial cooling, the Medieval Warm Period (MWP), centered around 1000 cal BP (Filion, 1984) and subsequent LIA cooling have been registered in northern Quebec ecosystems (Bhiry and Robert, 2006 ; Viau and Gajewski, 2009). Recurrent LIA cold/wet shifts have been observed in peatlands in northern Europe (De Vleeschouwer *et al.*, 2009 ; Mauquoy *et al.*, 2002b), coinciding with solar activity minima centered around AD 1300, 1500, 1700 and 1850. In Quebec, minimum temperatures and permafrost aggradation have been reported around 800 and 400-450 cal BP (Asselin and Payette, 2006 ;

Bhiry and Robert, 2006 ; Payette *et al.*, 1985 ; Viau and Gajewski, 2009) yet the minimum around AD 1800-1850 and the subsequent trend towards warmer and wetter conditions are better documented (Bergeron and Archambault, 1993 ; Bergeron *et al.*, 2001 ; Denneler *et al.*, 2008 ; Girardin *et al.*, 2004 ; Jacoby and D'Arrigo, 1989). Based on the number of standard deviations, our inferred dry shifts were more intense than wet shifts (Fig. 2.9), in spite of the apparent underestimation of WTD at the dry end of the gradient that is evident in cross validation of the transfer function (Booth, 2008).

Given the solid indications for long-term regional late-Holocene cooling and assuming that summer precipitation is more important to bog surface moisture than winter precipitation, the replicated dry surface events may be interpreted as forced by cold-dry climatic shifts. Indeed, the reconstructed dry shifts 8, 9 and 10, dated around 770, 400 and 140 cal BP, correspond to reported northern Quebec LIA cooling episodes. Previous research in LLC showed similar dry shifts between 1370-1070, 870-560 and 530-350 cal BP (Loisel and Garneau, 2010), while cooling-associated wet shifts were reconstructed in a peatland in Newfoundland around 600 and 200 cal BP (Hughes *et al.*, 2006). The 1370-1070 cal BP shift from Loisel and Garneau (2010) corresponds to dry shift 6 (Fig. 2.9), whereas the subsequent events coincide with shifts 8 and 9. Given the replicability and the extent of drying, we hypothesize that dry events have been amplified by a decadal persistence of frozen peat horizons. Some of the rapid shifts to near-surface water tables that followed these dry episodes (e.g. observed around 2400 cal BP in STE, 360 cal BP in LLC, as well as the ongoing rising water table in MOS), might thus be triggered by a collapse of frozen horizons. Melting effects may have been amplified by an autogenic effect: an anterior dry bog surface caused enhanced decomposition that concomitantly decreased hydraulic conductivity of the acrotelm peat, resulting in a potential for rapidly rising water tables and an extremely wet peatland surface. For instance, this autogenic effect may have amplified the effect of wetter climatic conditions causing high water tables (wet shift 11) since the end of the last LIA episode (Girardin *et al.*, 2006 ; Payette *et al.*, 2004).

Peatland thermokarst features are presently close to the Eastmain region, whereas the actual southern limit of sporadic peatland palsas is located ~140 km northward (Thibault and Payette, 2009). Although these residual palsas are not in equilibrium with present-day climate, past climatic variability might have been large enough to have caused episodic

frozen peatland surfaces in the Eastmain region. Nevertheless, except for testate amoeba-inferred extremes in WTD, associated dry-indicating vegetation assemblages and horizons of highly decomposed peat, we have no evidence for the past occurrence of persisting frozen peat in the Eastmain region. It is likely that LIA cooling resulted in declining C sequestration rates (Mauquoy *et al.*, 2002a), although high-resolution dating control is essential for confirmation.

2.6. Conclusion

Holocene peat C accumulation rates in three peatlands of the Eastmain region have been spatially and temporally highly variable, with Holocene averages between 14.9 and 22.6 g m^{-2} yr^{-1}. Rapid sequestration is associated with phases of important *Sphagnum* section *Acutifolia* presence during the warmer mid-Holocene, whereas low rates occur during the late-Holocene, when Cyperaceae and ligneous vegetation were abundant and water table levels showed high-amplitude fluctuations. However, high variability in C accumulation patterns implies a complex ecosystem response to driving factors and thus understanding of the interaction between climate, water table levels and vegetation mediated by local factors is essential in explaining differential tendencies. Varying patterns of C sequestration within the climatically uniform Eastmain region show that differential timing in ombrotrophication between coring sites may have been critical for patterns of long-term C sequestration.

Coinciding extremes in WTD may indicate the presence of seven cold/dry events in the region since 3000 cal BP. Possibly due to a more accurate chronology towards the present, three LIA cooling events may be best represented in the stratigraphy as indicated by (extremely) dry peatland surfaces. Given the past proximity of peatland palsas, episodes of persisting frozen peat might have contributed to the important water level drawdown in the studied peatlands.

Acknowledgements

We thank H. Asnong, P.-L. Dallaire and C. Vaillancourt for field assistance and A. Tadros, D. Charbonneau, N. Cliche Trudeau and M. Hayes for assistance in the laboratory. Thanks to the hydroclimatological scenarios team of the Ouranos consortium for providing climate data by manipulation of 1971-2003 NLWIS data. Funding was provided by Hydro-Quebec Production through the EM-1 Project *Reservoirs' net greenhouse gas emissions research project (2005-2009)*. We thank P. Hughes and an anonymous reviewer for comments on a previous draft. We appreciate the moments of discussion with *Les Tourbeux*.

References

Ali, A. A., H. Asselin, A. C. Larouche, Y. Bergeron, C. Carcaillet and P. J. H. Richard. 2008. «Changes in fire regime explain the Holocene rise and fall of Abies balsamea in the coniferous forests of western Quebec, Canada». *The Holocene*. vol. 18, no 5, p. 693-703.

Ali, A. A., C. Carcaillet and Y. Bergeron. 2009. «Long-term fire frequency variability in the eastern Canadian boreal forest: the influences of climate vs. local factors». *Global Change Biology*. vol. 15, no 5, p. 1230-1241.

Arseneault, D., and L. Sirois. 2004. «The millennial dynamics of a boreal forest stand from buried trees». *Journal of Ecology*. vol. 92, p. 490-504.

Asselin, H., and S. Payette. 2005. «Late Holocene opening of the forest tundra landscape in northern Québec, Canada». *Global Ecology and Biogeography*. vol. 14, no 4, p. 307-313.

Asselin, H., and S. Payette. 2006. «Origin and long-term dynamics of a subarctic tree line». *Ecoscience*. vol. 13, no 2, p. 135-142.

Ayotte, G., and L. Rochefort (2006). Les sphaignes du Québec. Québec, Université Laval: 136 p

Barber, K. E., F. M. Chambers, D. Maddy, R. Stoneman and J. S. Brew. 1994. «A sensitive high-resolution record of late Holocene climatic change from a raised bog in northern England». *The Holocene*. vol. 4, no 2, p. 198-205.

Beaulieu-Audy, V., M. Garneau, P. J. H. Richard and H. Asnong. 2009. «Holocene palaeoecological reconstruction of three boreal peatlands in the La Grande Riviere region, Quebec, Canada». *The Holocene*. vol. 19, no 3, p. 459-476.

Belyea, L. R., and R. S. Clymo. 2001. «Feedback control of the rate of peat formation». *Proceedings: Biological Sciences*. vol. 268, no 1473, p. 1315-1321.

Bergeron, Y., and S. Archambault. 1993. «Decreasing frequency of forest fires in the southern boreal zone of Quebec and its relation to global warming since the end of the 'Little Ice Age'». *The Holocene*. vol. 3, no 3, p. 255-259.

Bergeron, Y., S. Gauthier, V. Kafka, P. Lefort and D. Lesieur. 2001. «Natural fire frequency for the eastern Canadian boreal forest: consequences for sustainable forestry». *Canadian Journal of Forest Research*. vol. 31, p. 384-391.

Bhiry, N., S. Payette and É. C. Robert. 2007. «Peatland development at the arctic tree line (Québec, Canada) influenced by flooding and permafrost». *Quaternary Research*. vol. 67, no 3, p. 426-437.

Bhiry, N., and É. C. Robert. 2006. «Reconstruction of changes in vegetation and trophic conditions of a palsa in a permafrost peatland, subarctic Québec, Canada». *Ecoscience*. vol. 13, no 1, p. 56-65.

Blaauw, M., K. D. Bennett and J. A. Christen. 2010. «Random walk simulations of fossil proxy data». *The Holocene*. vol. 20, no 4, p. 645-649.

Booth, R. K. 2008. «Testate amoebae as proxies for mean annual water-table depth in Sphagnum-dominated peatlands of North America». *Journal of Quaternary Science*. vol. 23, no 1, p. 43-57.

Booth, R. K. 2010. «Testing the climate sensitivity of peat-based paleoclimate reconstructions in mid-continental North America». *Quaternary Science Reviews*. vol. 29, no 5-6, p. 720-731.

Booth, R. K., M. E. Sullivan and V. A. Sousa. 2008. «Ecology of testate amoebae in a North Carolina pocosin and their potential use as environmental and paleoenvironmental indicators». *Ecoscience*. vol. 15, no 2, p. 277-289.

Bubier, J. L., G. Bhatia, T. R. Moore, N. T. Roulet and P. M. Lafleur. 2003. «Spatial and temporal variability in growing-season net ecosystem carbon dioxide exchange at a large peatland in Ontario, Canada». *Ecosystems*. vol. 6, no 4, p. 353-367.

Carcaillet, C., and P. J. H. Richard. 2000. «Holocene changes in seasonal precipitation highlighted by fire incidence in eastern Canada». *Climate Dynamics*. vol. 16, no 7, p. 549-559.

Charman, D. J. 2002. *Peatlands and environmental change*. Chichester: John Wiley & Sons Ltd., 301 p.

Charman, D. J., K. E. Barber, M. Blaauw, P. G. Langdon, D. Mauquoy, T. J. Daley, P. D. M. Hughes and E. Karofeld. 2009. «Climate drivers for peatland palaeoclimate records». *Quaternary Science Reviews*. vol..28, no 19-20, p. 1811-1819.

Charman, D. J., A. Blundell, R. C. Chiverrell, D. Hendon and P. G. Langdon. 2006. «Compilation of non-annually resolved Holocene proxy climate records: stacked Holocene peatland palaeo-water table reconstructions from northern Britain». *Quaternary Science Reviews*. vol. 25, no 3-4, p. 336-350.

Charman, D. J., A. Blundell and ACCROTELM members. 2007. «A new European testate amoebae transfer function for palaeohydrological reconstruction on ombrotrophic peatlands». *Journal of Quaternary Science*. vol. 22, no 3, p. 209-221.

Charman, D. J., D. Hendon and S. Packman. 1999. «Multiproxy surface wetness records from replicate cores on an ombrotrophic mire: implications for Holocene palaeoclimate records». *Journal of Quaternary Science*. vol. 14, no 5, p. 451-463.

Charman, D. J., D. Hendon and W. A. Woodland. 2000. *The identification of testate amoebae (Protozoa: Rhizopoda) in peats*. Coll. «Technical Guide». London: Quaternary Research Association, 147 p.

Charman, D. J., and B. G. Warner. 1997. «The ecology of testate amoebae (Protozoa: Rhizopoda) in oceanic peatlands in Newfoundland, Canada: modelling hydrological

relationships for paleoenvironmental reconstruction». *Ecoscience*. vol. 4, no 4, p. 555-562.

Clymo, R. S. 1984. «The limits to peat bog growth». *Philosophical Transactions of the Royal Society of London. Series B, Biological Sciences*. vol. 303, no 1117, p. 605-654.

De Vleeschouwer, F., N. Piotrowska, J. Sikorski, J. Pawlyta, A. Cheburkin, G. Le Roux, M. Lamentowicz, N. Fagel and D. Mauquoy. 2009. «Multiproxy evidence of 'Little Ice Age' palaeoenvironmental changes in a peat bog from northern Poland». *The Holocene*. vol. 19, no 4, p. 625-637.

Dean, W. E. 1974. «Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition; comparison with other methods». *Journal of Sedimentary Research*. vol. 44, no 1, p. 242-248.

Denneler, B., H. Asselin, Y. Bergeron and Y. Bégin. 2008. «Decreased fire frequency and increased water levels affect riparian forest dynamics in southwestern boreal Quebec, Canada». *Canadian Journal of Forest Research*. vol. 38, no 5, p. 1083-1094.

Eppinga, M. B., M. Rietkerk, M. J. Wassen and P. C. De Ruiter. 2009. «Linking habitat modification to catastrophic shifts and vegetation patterns in bogs». *Plant Ecology*. vol. 200, p. 53-68.

Filion, L. 1984. «A relationship between dunes, fire and climate recorded in the Holocene deposits of Quebec». *Nature*. vol. 309, no 5968, p. 543-546.

Frolking, S., and N. T. Roulet. 2007. «Holocene radiative forcing impact of northern peatland carbon accumulation and methane emissions». *Global Change Biology*. vol. 13, no 5, p. 1079-1088.

Garneau, M. (1995). Collection de référence de graines et autres macrofossiles végétaux de taxons provenant du Québec méridional et boréal et de l'arctique canadien. Division de la science des terrains. Sainte-Foy, Geological Survey of Canada: 47 p

Girardin, M. P., J. Tardif, M. D. Flannigan and Y. Bergeron. 2004. «Multicentury reconstruction of the Canadian Drought Code from eastern Canada and its relationship with paleoclimatic indices of atmospheric circulation». *Climate Dynamics*. vol. 23, no 2, p. 99-115.

Girardin, M. P., J. C. Tardif, M. D. Flannigan and Y. Bergeron. 2006. «Synoptic-scale atmospheric circulation and boreal Canada summer drought variability of the past three centuries». *Journal of Climate*. vol. 19, no 10, p. 1922-1947.

Glaser, P. H., B. Hansen, C. S., D. I. Siegel, A. S. Reeve and P. J. Morin. 2004. «Rates, pathways and drivers for peatland development in the Hudson Bay Lowlands, northern Ontario, Canada». *Journal of Ecology*. vol. 92, no 6, p. 1036-1053.

Haslett, J., and A. Parnell. 2008. «A simple monotone process with application to radiocarbon-dated depth chronologies». *Journal of the Royal Statistical Society: Series C*. vol. 57, no 4, p. 399-418.

Heijmans, M. M. P. D., F. Berendse, W. J. Arp, A. K. Masselink, H. Klees, W. de Visser and N. van Breemen. 2001. «Effects of elevated carbon dioxide and increased nitrogen deposition on bog vegetation in the Netherlands». *Journal of Ecology*. vol. 89, no 2, p. 268-279.

Heiri, O., A. F. Lotter and G. Lemcke. 2001. «Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results». *Journal of Paleolimnology*. vol. 25, no 1, p. 101-110.

Hendon, D., and D. J. Charman. 1997. «The preparation of testate amoebae (Protozoa: Rhizopoda) samples from peat». *The Holocene*. vol. 7, no 2, p. 199-205.

Hughes, P. D. M., A. Blundell, D. J. Charman, S. Bartlett, J. R. G. Daniell, A. Wojatschke and F. M. Chambers. 2006. «An 8500 cal. year multi-proxy climate record from a bog in eastern Newfoundland: contributions of meltwater discharge and solar forcing». *Quaternary Science Reviews*. vol. 25, no 11-12, p. 1208-1227.

Hutchinson, M. F., D. W. McKenney, K. Lawrence, J. H. Pedlar, R. F. Hopkinson, E. Milewska and P. Papadopol. 2009. «Development and testing of Canada-wide interpolated spatial models of daily minimum–maximum temperature and precipitation for 1961–2003». *Journal of Applied Meteorology and Climatology*. vol. 48, p. 725-741.

Jacoby, G. C., and R. D'Arrigo. 1989. «Reconstructed Northern Hemisphere annual temperature since 1671 based on high-latitude tree-ring data from North America». *Climatic Change*. vol. 14, no 1, p. 39-59.

Juggins, S. (2007). C2 Version 1.5 User guide. Software for ecological and palaeoecological data analysis and visualisation. Newcastle upon Tyne, Newcastle University

Kaufman, D. S., T. A. Ager, N. J. Anderson, P. M. Anderson, J. T. Andrews, P. J. Bartlein, L. B. Brubaker, L. L. Coats, L. C. Cwynar, M. L. Duvall, A. S. Dyke, M. E. Edwards, W. R. Eisner, K. Gajewski, A. Geirsdóttir, F. S. Hu, A. E. Jennings, M. R. Kaplan, M. W. Kerwin, A. V. Lozhkin, G. M. MacDonald, G. H. Miller, C. J. Mock, W. W. Oswald, B. L. Otto-Bliesner, D. F. Porinchu, K. Rühland, J. P. Smol, E. J. Steig and B. B. Wolfe. 2004. «Holocene thermal maximum in the western Arctic (0-180°W)». *Quaternary Science Reviews*. vol. 23, no 5-6, p. 529-560.

Kerwin, M. W., J. T. Overpeck, R. S. Webb and K. H. Anderson. 2004. «Pollen-based summer temperature reconstructions for the eastern Canadian boreal forest, subarctic, and Arctic». *Quaternary Science Reviews*. vol. 23, no 18-19, p. 1901-1924.

Kuhry, P. 1994. «The role of fire in the development of Sphagnum-dominated peatlands in western boreal Canada». *Journal of Ecology*. vol. 82, no 4, p. 899-910.

Lamarre, A. 2010. «Développement d'une fonction de transfert au moyen des thécamoebiens et reconstitution des conditions paléoenvironnementales holocènes d'une tourbière à palses, Kuujjuarapik, Québec nordique». Montreal, Département de Géographie, Université du Québec à Montréal, 116 p.

Loisel, J., and M. Garneau. 2010. «Late Holocene paleoecohydrology and carbon accumulation estimates from two boreal peat bogs in eastern Canada: Potential and limits of multi-proxy archives». *Palaeogeography, Palaeoclimatology, Palaeoecology*. vol. 291, no 3-4, p. 493-533.

Marie-Victorin, F. 1995. *Flore Laurentienne, third edition*: Les Presses de l'Université de Montréal, 1112 p.

Mauquoy, D., T. Engelkes, M. H. M. Groot, F. Markesteijn, M. G. Oudejans, J. van der Plicht and B. van Geel. 2002a. «High-resolution records of late-Holocene climate change and carbon accumulation in two north-west European ombrotrophic peat bogs». *Palaeogeography, Palaeoclimatology, Palaeoecology*. vol. 186, no 3-4, p. 275-310.

Mauquoy, D., B. van Geel, M. Blaauw and J. van der Plicht. 2002b. «Evidence from northwest European bogs shows 'Little Ice Age' climatic changes driven by variations in solar activity». *The Holocene*. vol. 12, no 1, p. 1-6.

Miousse, L., N. Bhiry and M. Lavoie. 2003. «Isolation and water-level fluctuations of Lake Kachishayoot, Northern Québec, Canada». *Quaternary Research*. vol. 60, no 2, p. 149-161.

Mitchell, E., D. Charman and B. Warner. 2008. «Testate amoebae analysis in ecological and paleoecological studies of wetlands: past, present and future». *Biodiversity and Conservation*. vol. 17, no 9, p. 2115-2137.

Nilsson, M., M. Klarqvist, E. Bohlin and G. Possnert. 2001. «Variation in ^{14}C age of macrofossils and different fractions of minute peat samples dated by AMS». *The Holocene*. vol. 11, no 5, p. 579-586.

Payette, S., A. Delwaide, M. Caccianiga and M. Beauchemin. 2004. «Accelerated thawing of subarctic peatland permafrost over the last 50 years». *Geophysical Research Letters*. vol. 31, no 18, p. L18208.

Payette, S., and L. Filion. 1993. «Holocene water-level fluctuations of a subarctic lake at the tree line in northern Québec». *Boreas*. vol. 22, no 1, p. 7-14.

Payette, S., L. Filion, L. Gauthier and Y. Boutin. 1985. «Secular climate change in old-growth tree-line vegetation of northern Quebec». *Nature*. vol. 315, no 6015, p. 135-138.

Payette, S., and R. Gagnon. 1985. «Late Holocene deforestation and tree regeneration in the forest-tundra of Quebec». *Nature*. vol. 313, no 6003, p. 570-572.

Payne, R. J., and E. A. D. Mitchell. 2008. «How many is enough? Determining optimal count totals for ecological and palaeoecological studies of testate amoebae». *Journal of Paleolimnology*, p. 1-13.

Pelletier, L., M. Garneau and T. R. Moore. in review. «Interannual variation in net ecosystem exchange at the microform scale in a boreal bog, Eastmain region, Quebec, Canada». *Journal of Geophysical Research - Biogeosciences*.

Pitkänen, A., J. Turunen and K. Tolonen. 1999. «The role of fire in the carbon dynamics of a mire, eastern Finland». *The Holocene*. vol. 9, no 4, p. 453-462.

R Development Core Team. 2009. R: A language and environment for statistical computing. Vienna, R Foundation for Statistical Computing.

Reimer, P. J., M. G. L. Baillie, E. Bard, A. Bayliss, J. W. Beck, C. J. H. Bertrand, P. G. Blackwell, C. E. Buck, G. S. Burr, K. B. Cutler, P. E. Damon, R. L. Edwards, R. G. Fairbanks, M. Friedrich, T. P. Guilderson, A. G. Hogg, K. A. Hughen, B. Kromer, G. McCormac, S. Manning, C. Bronk Ramsey, R. W. Reimer, S. Remmele, J. R. Southon, M. Stuiver, S. Talamo, F. W. Taylor, J. van der Plicht and C. E. Weyhenmeyer. 2004. «INTCAL04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP». *Radiocarbon*. vol. 46, no 3, p. 1029-1058.

Silvola, J., J. Alm, U. Ahlholm, H. Nykanen and P. J. Martikainen. 1996. «CO_2 fluxes from peat in boreal mires under varying temperature and moisture conditions». *Journal of Ecology*. vol. 84, no 2, p. 219-228.

Sullivan, M. E., and R. K. Booth. in review. «The influence of short-term environmental variability on the composition of testate amoeba communities in Sphagnum peatlands». *Microbial Ecology*.

Swanson, D. K. 2007. «Interaction of mire microtopography, water supply, and peat accumulation in boreal mires». *Suo*. vol. 58, no 2, p. 37-47.

Thibault, S., and S. Payette. 2009. «Recent permafrost degradation in bogs of the James Bay area, northern Quebec, Canada». *Permafrost and Periglacial Processes*. vol. 20, no 4, p. 383-389.

Tolonen, K., and J. Turunen. 1996. «Accumulation rates of carbon in mires in Finland and implications for climate change». *The Holocene*. vol. 6, no 2, p. 171-178.

Turetsky, M. R., R. K. Wieder, D. H. Vitt, R. J. Evans and K. D. Scott. 2007. «The disappearance of relict permafrost in boreal north America: effects on peatland carbon storage and fluxes». *Global Change Biology*. vol. 13, p. 1922-1934.

Turunen, C., and J. Turunen. 2003. «Development history and carbon accumulation of a slope bog in oceanic British Columbia, Canada». *The Holocene*. vol. 13, no 2, p. 225-238.

Turunen, J., T. Tahvanainen, K. Tolonen and A. Pitkänen. 2001. «Carbon accumulation in West Siberian mires, Russia». *Global Biogeochemical Cycles*. vol. 15, no 2, p. 285-296.

Turunen, J., E. Tomppo, K. Tolonen and A. Reinikainen. 2002. «Estimating carbon accumulation rates of undrained mires in Finland-application to boreal and subarctic regions». *The Holocene*. vol. 12, no 1, p. 69-80.

van Bellen, S., P.-L. Dallaire, M. Garneau and Y. Bergeron. 2011. «Quantifying spatial and temporal Holocene carbon accumulation in ombrotrophic peatlands of the Eastmain region, Quebec, Canada». *Global Biogeochemical Cycles*.

Viau, A. E., and K. Gajewski. 2009. «Reconstructing millennial-scale, regional paleoclimates of boreal Canada during the Holocene». *Journal of Climate*. vol. 22, no 2, p. 316-330.

Viau, A. E., K. Gajewski, M. C. Sawada and P. Fines. 2006. «Millennial-scale temperature variations in North America during the Holocene». *Journal of Geophysical Research*. vol. 111.

Warner, B., T. Asada and N. Quinn. 2007. «Seasonal influences on the ecology of testate amoebae (Protozoa) in a small Sphagnum peatland in southern Ontario, Canada». *Microbial Ecology*. vol. 54, no 1, p. 91-100.

Yu, Z. 2006a. «Holocene carbon accumulation of fen peatlands in boreal western Canada: a complex ecosystem response to climate variation and disturbance». *Ecosystems*. vol. 9, no 8, p. 1278-1288.

Yu, Z. 2006b. «Power laws governing hydrology and carbon dynamics in northern peatlands». *Global and Planetary Change*. vol. 53, no 3, p. 169-175.

Yu, Z., D. W. Beilman and M. C. Jones. 2009. «Sensitivity of northern peatland carbon dynamics to Holocene climate change». In *Carbon cycling in northern peatlands,* Andrew J. Baird, Lisa R. Belyea, Xavier Comas, A. S. Reeve et Lee D. Slater, p. 55-69. Washington: American Geophysical Union.

Yu, Z., J. Loisel, D. P. Brosseau, D. W. Beilman and S. J. Hunt. 2010. «Global peatland dynamics since the Last Glacial Maximum». *Geophysical Research Letters*. vol. 37, no 13, p. L13402.

Yu, Z., D. H. Vitt, I. D. Campbell and M. J. Apps. 2003. «Understanding Holocene peat accumulation pattern of continental fens in western Canada». *Canadian Journal of Botany*. vol. 81, p. 267-282.

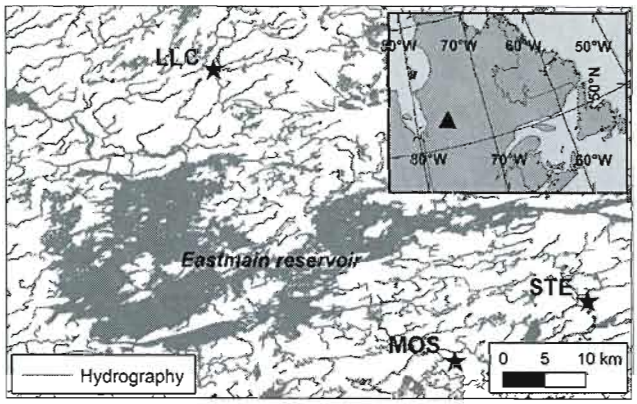

Figure 2.1. Study region.

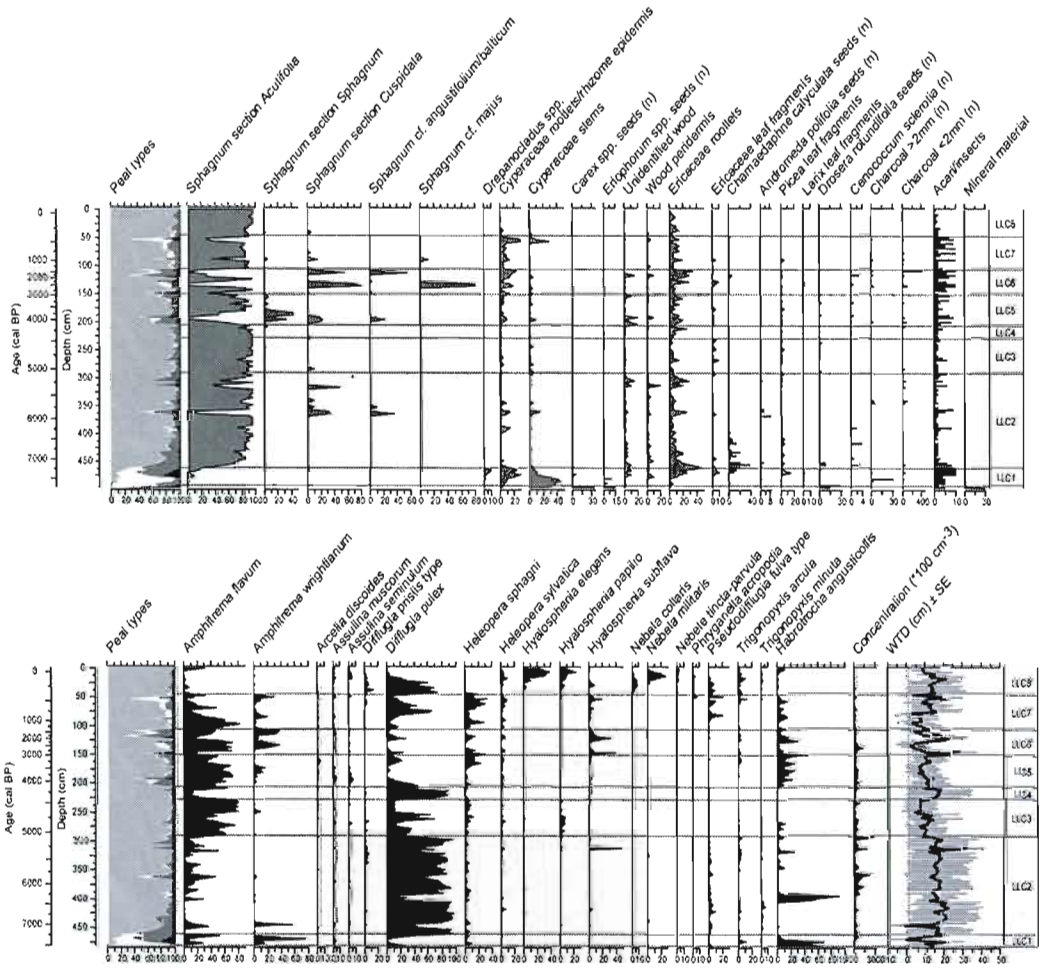

Figure 2.2. Plant macrofossil (upper) and testate amoeba (lower) diagram for LLC. Plant macrofossils are expressed as volume percentages or absolute quantities (n); amoeba presence is expressed as a percentage. Peat types are subdivided into *Sphagnum* peat (light grey), sedge peat (white), ligneous peat (dark grey) and unidentified organic matter (black). Inferred WTD (±SE) is shown at the right of the testate amoeba assemblages.

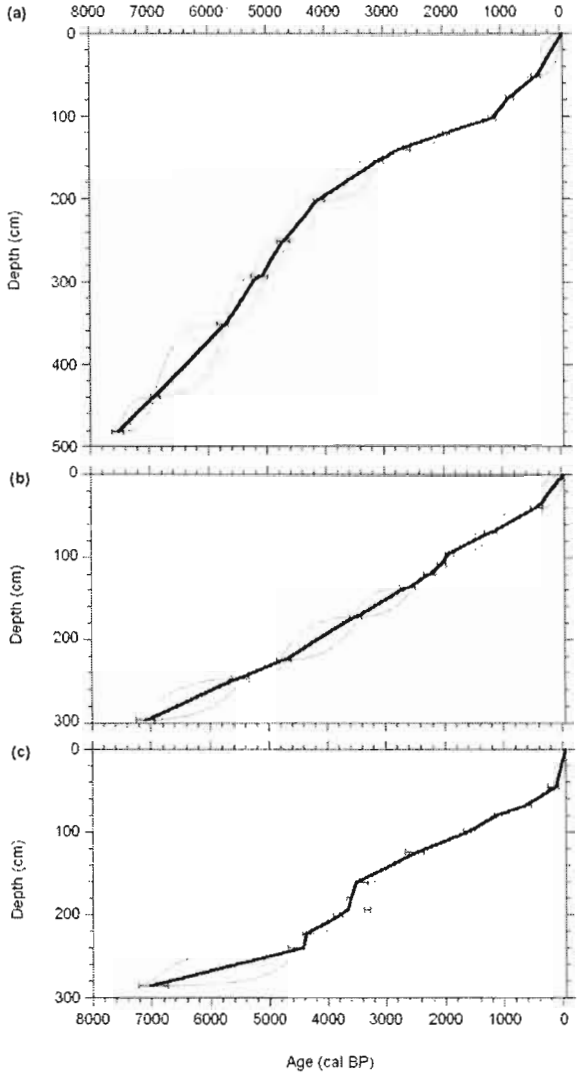

Figure 2.3. Age-depth models (solid black line). Grey lines mark the 95% confidence interval. a) LLC; b) MOS; c) STE.

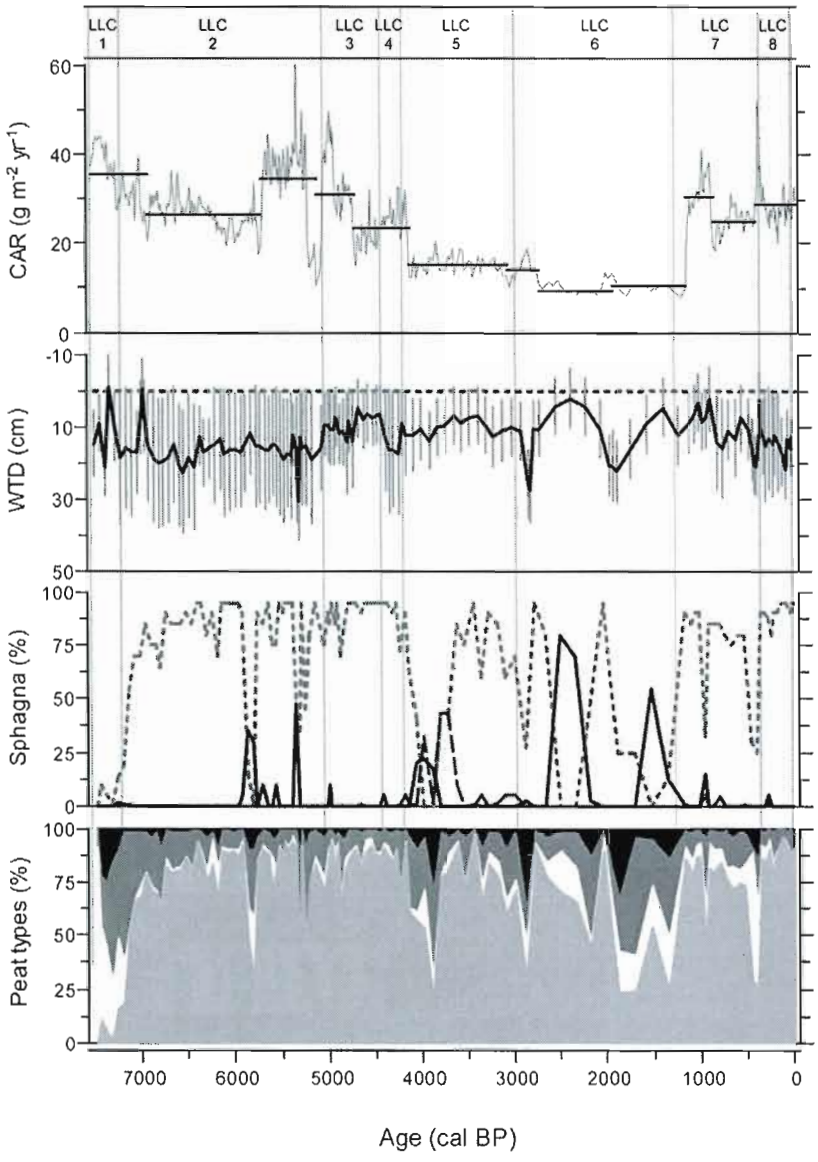

Figure 2.4. LLC C accumulation rate (CAR), WTD (±SE), *Sphagna* presence and peat types. CAR are shown at high resolution (grey) and as mean values between dated levels (black). *Sphagna* shows *Sphagnum* sections *Acutifolia* (dotted line), *Sphagnum* (dashed line) and *Cuspidata* (solid line). Peat types are subdivided into *Sphagnum* peat (light grey), sedge peat (white), ligneous peat (dark grey) and unidentified organic matter (black).

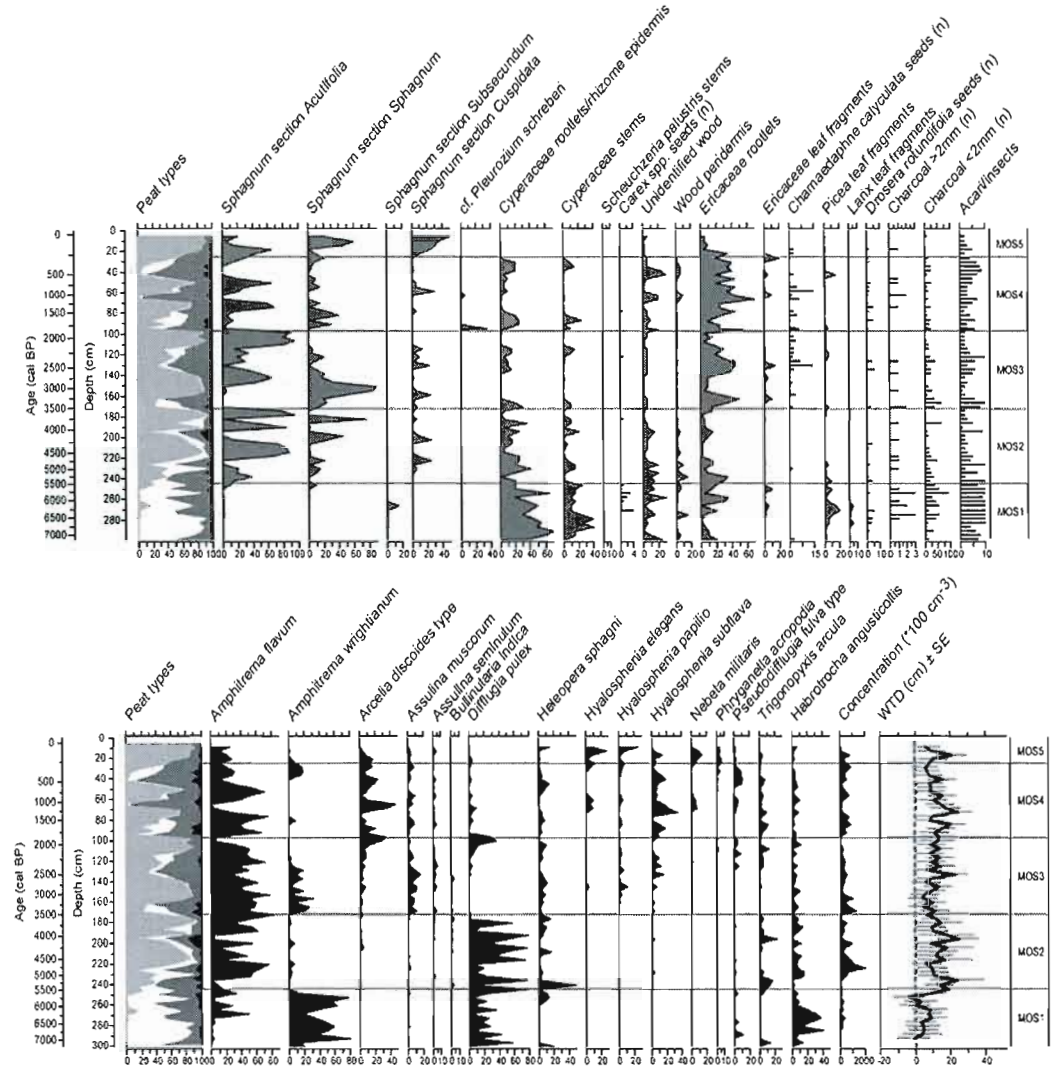

Figure 2.5. Plant macrofossil (upper) and testate amoeba (lower) diagram for MOS. Plant macrofossils are expressed as volume percentages or absolute quantities (n); amoeba presence is expressed as a percentage. Peat types are subdivided into *Sphagnum* peat (light grey), sedge peat (white), ligneous peat (dark grey) and unidentified organic matter (black). Inferred WTD (±SE) is shown at the right of the testate amoeba assemblages.

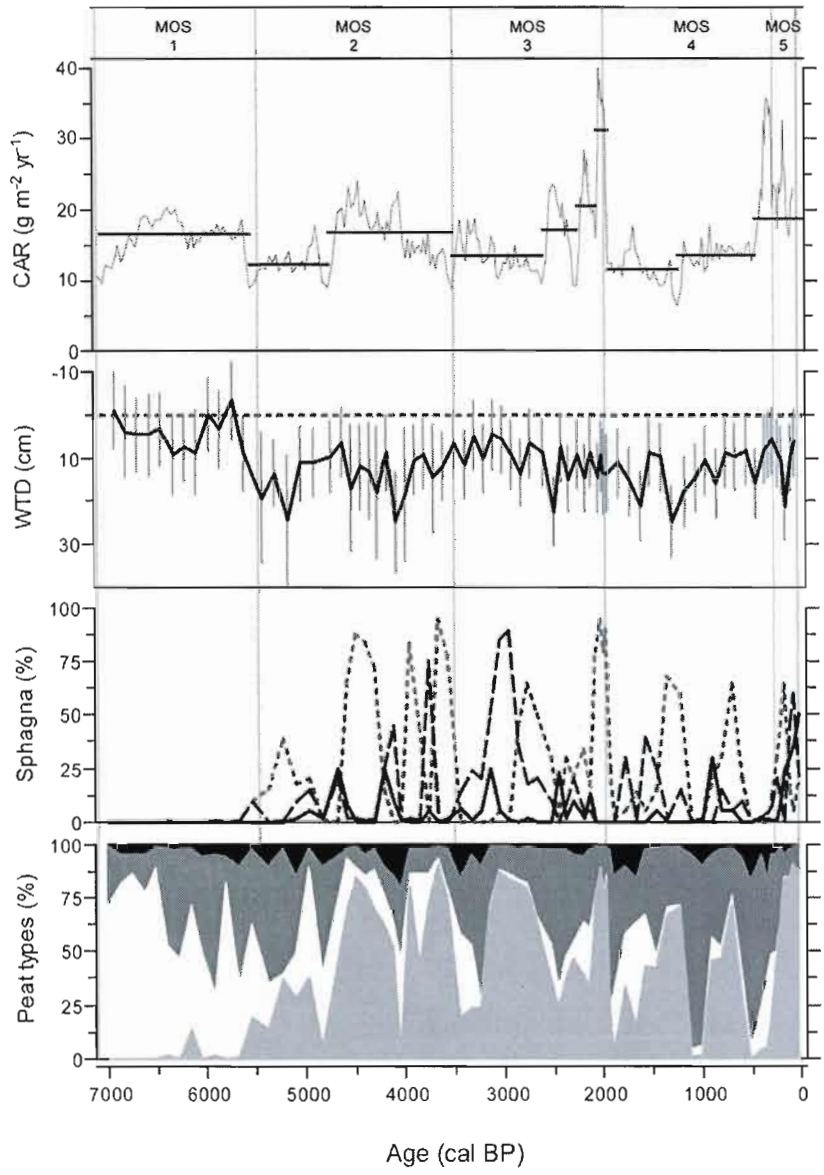

Figure 2.6. MOS C accumulation rate (CAR), WTD (±SE), *Sphagna* presence and peat types. CAR are shown at high resolution (grey) and as mean values between dated levels (black). *Sphagna* shows *Sphagnum* sections *Acutifolia* (dotted line), *Sphagnum* (dashed line) and *Cuspidata* (solid line). Peat types are subdivided into *Sphagnum* peat (light grey), sedge peat (white), ligneous peat (dark grey) and unidentified organic matter (black).

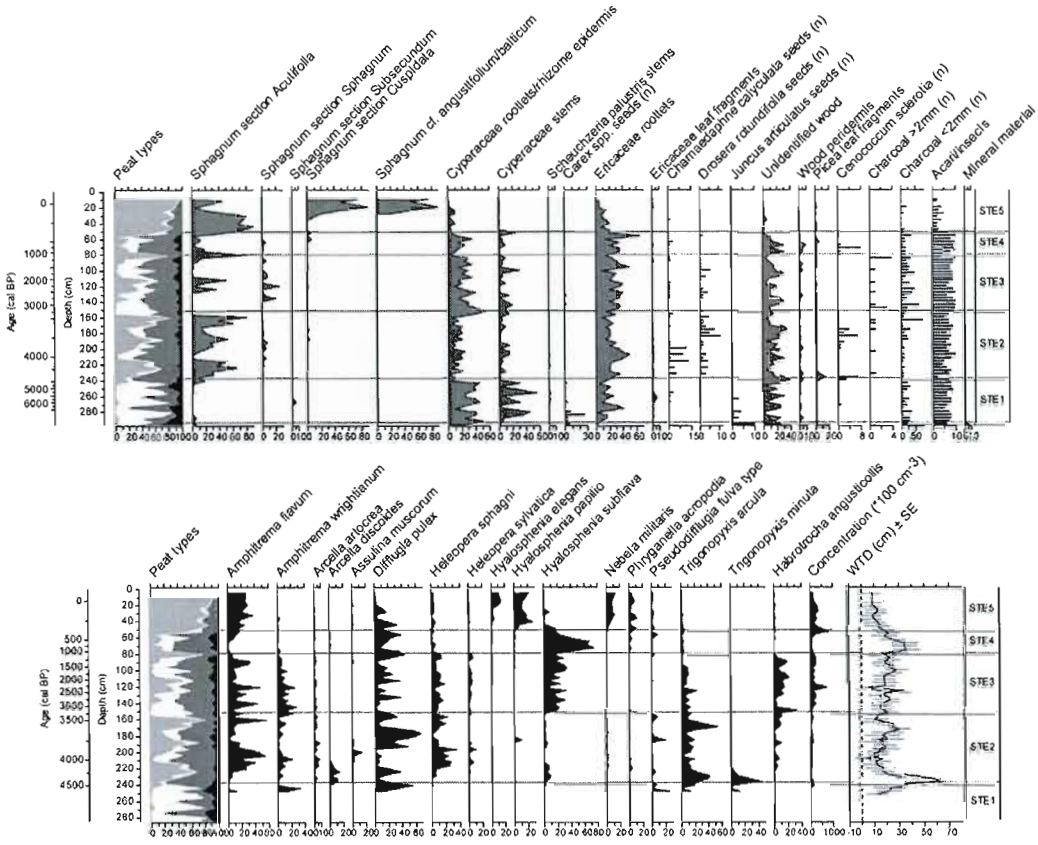

Figure 2.7. Plant macrofossil (upper) and testate amoeba (lower) diagram for STE. Plant macrofossils are expressed as volume percentages or absolute quantities (n); amoeba presence is expressed as a percentage. Peat types are subdivided into *Sphagnum* peat (light grey), sedge peat (white), ligneous peat (dark grey) and unidentified organic matter (black). Inferred WTD (±SE) is shown at the right of the testate amoeba assemblages.

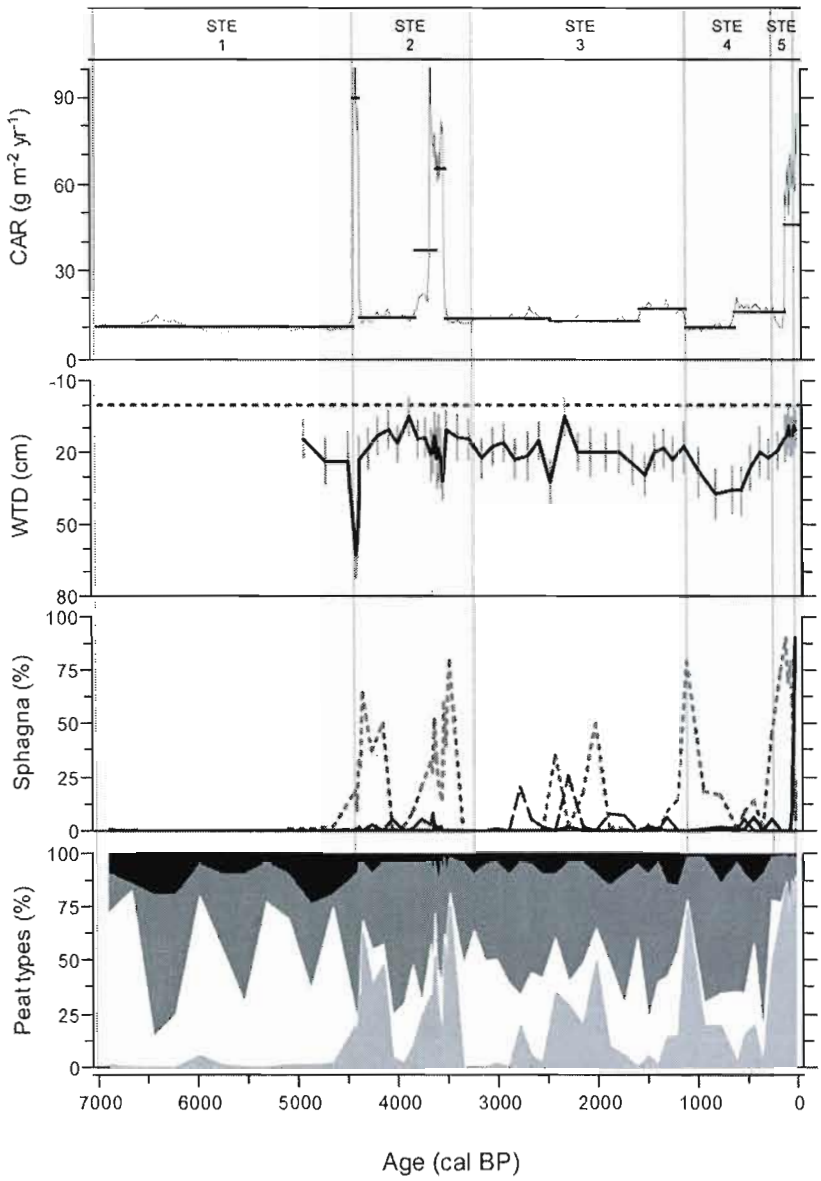

Figure 2.8. STE C accumulation rate (CAR), WTD (±SE), *Sphagna* presence and peat types. CAR are shown at high resolution (grey) and as mean values between dated levels (black). *Sphagna* shows *Sphagnum* sections *Acutifolia* (dotted line), *Sphagnum* (dashed line) and *Cuspidata* (solid line). Peat types are subdivided into *Sphagnum* peat (light grey), sedge peat (white), ligneous peat (dark grey) and unidentified organic matter (black).

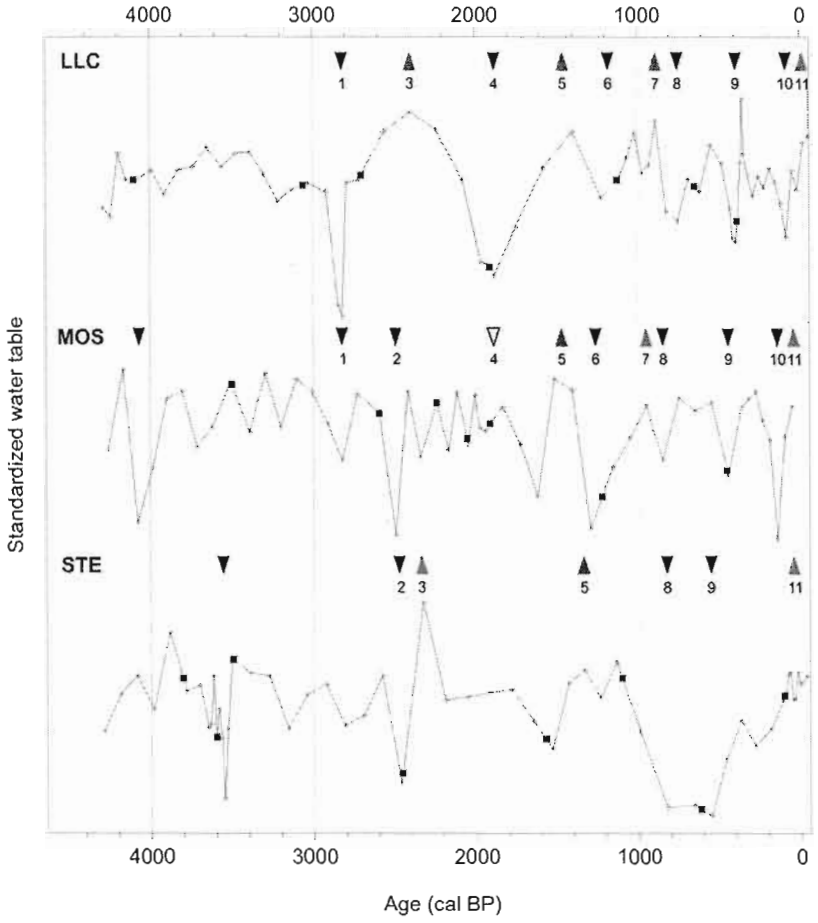

Figure 2.9. Detrended and normalized late-Holocene WTD reconstructions. Black triangles indicate dry shifts, grey triangles indicate wet shifts and open triangle indicates a dry shift that was observed from plant macrofossils only. Black squares show dated levels. Equally numbered shifts may be corresponding taking into account the 95% confidence interval for the level considered.

Site	Core	Sample depth (cm)	Laboratory number	Material dated	^{14}C age (BP)	2σ range (cal BP)	Age (cal BP)
LLC	C	51-52	UCIAMS43480	Sph stems	340±20	317-480	391
	C	77-78	UCIAMS58634	Sph stems	915±15	777-906	846
	C	102-103	UCIAMS50203	Sph stems	1205±15	1072-1225	1140
	C	120-121	UCIAMS57419	Sph stems	1980±15	1872-1991	1924
	C	140-141	UCIAMS57421	Sph stems	2550±15	2531-2748	2716
	C	153-154	UCIAMS50204	Sph stems	2915±15	2984-3155	3057
	C	201-202	UCIAMS43479	Sph stems	3745±20	3996-4171	4107
	C	250-251	UCIAMS58636	Sph stems	4165±20	4596-4819	4701
	C	293-294	UCIAMS50205	Sph stems	4450±15	4980-5265	5110
	C	351-352	UCIAMS43478	Sph stems	4985±20	5653-5844	5701
	C	439-440	UCIAMS50206	Sph stems	6055±15	6821-6968	6912
	C	480-483	Beta223743	Eric leaf frs	6640±40	7431-7627	7520
MOS	C	40-41	UCIAMS57424	Sph stems	355±15	325-522	444
	C	70-71	UCIAMS54958	Sph stems	1270±25	1095-1313	1223
	C	95-96	UCIAMS64586	Sph stems	1990±20	1837-1985	1924
	C	108-109	UCIAMS67515	Sph stems	2065±25	1976-2119	2043
	C	120-121	UCIAMS54959	Sph stems	2225±25	2157-2335	2237
	C	136-137	UCIAMS64588	Sph stems	2490±20	2478-2728	2591
	C	172-173	UCIAMS54960	Sph stems	3275±25	3405-3604	3506
	C	224-225	UCIAMS54961	Sph stems	4185±25	4609-4863	4739
	C	246-247	UCIAMS57426	Sph stems	4740±15	5333-5634	5534
	C	296-297	Beta223744	Sph stems	6200±40	6936-7237	7072
STE	C	45-46	UCIAMS54962	Sph stems	105±30	75-260	116
	C	67-68	UCIAMS58645	Picea leaf frs	600±20	548-646	620
	C	79-80	UCIAMS64589	Sph stems	1175±20	1039-1166	1112
	C	98-99	UCIAMS54963	Sph stems	1715±25	1568-1697	1578
	C	124-125	UCIAMS65381	Sph stems	2445±20	2351-2663	2461
	C	160-161	UCIAMS54964	Sph stems	3255±30	3316-3521	3505
	C	179-180	UCIAMS67514	Sph stems	3415±25	3588-3671	3597
	C	201-202	UCIAMS65382	Sph stems	3485±20	3757-3912	3805
	C	223-224	UCIAMS54965	Sph stems	3960±30	4283-4431	4362
	C	239-240	UCIAMS58643	Sph stems/Picea leaf frs	3975±15	4422-4685	4424
	C	285-286	UCIAMS40360	Sph stems	6225±20	6731-7219	7007

Table 2.1. Radiocarbon datings sorted by core.

Zone	Depth (cm)	Age (cal BP)	Main features
LLC8	45-0	330-present	Well- preserved *Sphagnum* section *Acutifolia* peat with *Difflugia pulex* and water tables between 9-21 cm below the surface with an important dry shift around 100 cal BP.
LLC7	105-45	1250-330	Sphagnum peat with both *Amphitrema flavum* and *Difflugia pulex*. Fluctuating water tables with persisting lowest water tables around 410 cal BP, associated with episodic highly decomposed peat
LLC6	150-105	2960-1250	Decomposed *Sphagnum* peat with Ericaceae and Cyperaceae. Highly fluctuating water tables (2-30 cm below surface) with rapid alternation of wet- and dry-indicating amoebae. Deep water tables (~22-27 cm below the surface) centered ~1920 and ~2830 cal BP
LLC5	206-150	4170-2960	Decomposed *Sphagnum* peat with occasional Ericaceae and Cyperaceae. Stable water tables (7-13 cm below surface) with *Amphitrema flavum* and *Habrotrocha angusticollis*
LLC4	228-206	4420-4170	Well-preserved *Sphagnum* section *Acutifolia* peat with *Difflugia pulex* and stable water tables (15-20 cm below surface)
LLC3	290-228	5040-4420	Well preserved *Sphagnum* section *Acutifolia* peat with *Amphitrema flavum* and high water tables (5-15 cm below surface)
LLC2	460-290	7210-5040	Well-preserved *Sphagnum* section *Acutifolia* peat with *Difflugia pulex* and stable water tables (15-20 cm below surface)
LLC1	482-460	7520-7210	Wet fen conditions: decomposed Cyperaceae and *Drepanocladus* peat with *Habrotrocha angusticollis* and *Amphitrema wrightianum*

Table 2.2. Stratigraphic zonation for LLC.

Zone	Depth (cm)	Age (cal BP)	Main features
MOS5	25-0	240-present	Well-preserved *Sphagnum* peat with a single dry shift at 17 cm/150 cal BP
MOS4	97-25	1940-240	Ericaceae-*Sphagnum* peat with *Amphitrema flavum*, *Arcella discoides* and *Hyalosphenia subflava*. Fluctuating water tables (6-25 cm below surface) with important dry shifts centered around 1890 cal BP (presence of *Pleurozium schreberi*) and 1290 cal BP (presence of *Hyalosphenia subflava* and *Trigonopyxis arcula*).
MOS3	172-97	3480-1940	*Sphagnum*-Ericaceae peat with *Amphitrema flavum*. Relatively stable water tables and accelerating accumulation toward the end of the period
MOS2	245-172	5450-3480	*Sphagnum*-Cyperaceae peat with *Difflugia pulex* and *Amphitrema flavum*. Fluctuating water tables (6-25 cm below surface)
MOS1	297-245	7070-5450	Wet fen conditions. Cyperaceae and Ericaceae peat with *Amphitrema wrightianum*, *Difflugia pulex* and *Habrotrocha angusticollis*

Table 2.3. Stratigraphic zonation for MOS.

Zone	Depth (cm)	Age (cal BP)	Main features
STE5	50-0	220-present	Well-preserved *Sphagnum* peat with *Amphitrema flavum and Hyalosphenia papilio*, shifting toward *Cuspidata* section dominance
STE4	79-50	1090-220	Highly decomposed Ericaceae/wood-Cyperaceae peat with *Hyalosphenia subflava* and extremely low water tables (20-37 cm below surface)
STE3	151-79	3230-1090	Decomposed Ericaceae/wood-Cyperaceae peat with *Sphagnum* and *Hyalosphenia subflava*. Unstable water tables.
STE2	237-151	4410-3230	*Ericaceae*/wood-*Sphagnum* peat with *Amphitrema flavum*, *Difflugia pulex* and *Trigonopyxis arcula*. Highly fluctuating water tables (5-31 cm below surface)
STE1	286-237	7010-4410	Fen conditions. Highly decomposed Cyperaceae and Ericaceae peat

Table 2.4. Stratigraphic zonation for STE.

CHAPTER III

DID FIRES DRIVE HOLOCENE CARBON SEQUESTRATION IN BOREAL OMBROTROPHIC PEATLANDS OF EASTERN CANADA?

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Article submitted to Quaternary Research.

Résumé

Les feux constituent un facteur important en ce qui concerne l'accumulation du carbone dans les écosystèmes boréaux de l'Amérique du Nord. Dans l'ensemble de la région boréale, la dynamique des feux de tourbière semble être différente de celle des peuplements forestiers, surtout dans les écosystèmes tourbeux ouverts et relativement humides trouvés fréquemment au nord-est du Canada. L'objectif de la recherche était de déterminer l'impact des feux de tourbière sur l'accumulation du carbone durant l'Holocène, et de vérifier si la végétation des tourbières a influencé les évènements de feu. Afin de couvrir la variabilité spatiale, deux carottes par tourbière ont été prélevées dans trois tourbières de la région d'Eastmain afin de procéder à la quantification de leur contenu en charbon.

Les résultats montrent que les régimes de feux ont été très variables à l'échelle spatiale et temporelle, alors qu'une tendance régionale de ralentissement de l'accumulation du carbone a été identifiée autour de 2000 cal BP. L'absence d'une corrélation négative entre les régimes des feux de tourbière et l'accumulation du carbone indique que ceux-ci n'ont pas exercé d'influence déterminante dans l'accumulation du carbone durant l'Holocène. Puisque les assemblages végétaux ne concordaient pas non plus aux évènements de feu, on peut conclure que ces derniers ont plutôt été favorisés par des épisodes d'assèchement de la surface de la tourbière au cours de quelques années ou décennies, une période trop courte pour être détectée par macrorestes végétaux. En plus, les conditions hydrologiques hétérogènes des tourbières compliquent l'identification des conditions propices aux feux de tourbière.

Abstract

Wildfire is an important factor considering carbon sequestration in the North American boreal biomes. Within this biome, peatland fire dynamics may differ from those in upland forests, especially in open, wet ombrotrophic ecosystems as found in northeastern Canada. We aimed to determine to which extent peat fire events have influenced carbon accumulation during the Holocene, and to verify if local vegetation has influenced the fire regimes. To cover spatial variability, two cores per peatland were extracted from three peatlands in the Eastmain region and analyzed for stratigraphic charcoal accumulation.

Regional peatland fire patterns showed high spatial and temporal variability, while carbon accumulation rates generally declined towards the late-Holocene. The absence of a negative correlation between peatland fire regimes and carbon accumulation indicates that fire regimes have not been a driving factor considering Holocene carbon sequestration. As trends in the reconstructed vegetation records did not match with fire events either, we conclude that fire occurrence may be rather linked to annual to subdecadal shifts in local hydrological conditions, relatively independent of vegetation, that are not detectable by stratigraphic plant macrofossil analyses. Moreover, heterogenous hydrological conditions complicate the identification of factors determining peatland burning potential.

3.1. Introduction

Wildfire and climate generally determine patterns of plant communities and species distribution in the Canadian boreal region (Payette, 1992). Fire frequency and severity (i.e. depth of forest floor consumption) are highly variable within the Canadian boreal forest resulting from variations in climate, topography, stand density and species effects. Nevertheless, high-intensity crown fire is a dominant fire type in the upland forest ecosystem (Conny and Slater, 2002).

Within the boreal and subarctic regions peatlands are frequent in topographic depressions, globally covering 3-4 million km^2 (MacDonald *et al.*, 2006 ; Yu *et al.*, 2010). Due to cold, humid, nutrient-poor and acidic conditions that restrict decomposition, peatlands sequester organic carbon (C) and expanded laterally and vertically over millennia (Korhola, 1994 ; Korhola *et al.*, 2010 ; Tolonen and Turunen, 1996). As a result, the global northern peatland C stock, that started accumulating after the Last Glacial Maximum, presently attains ~547 (range: 473-621) Pg (Yu *et al.*, 2010), constituting approximately a third of global soil C.

Peatland hydrology and vegetation strongly influence the potential for fire propagation, while the amount of peat consumed varies with local and regional humidity conditions (Zoltai *et al.*, 1998). The highly variable spatial patterns of moisture conditions within and among peatlands are closely linked to surface microtopography (i.e. hummocks and hollows) and vegetation composition. Hence, interactions between vegetation and hydrology are important factors considering burning potential (Higuera *et al.*, 2009). The presence of trees and shrubs in peatlands positively influences fuel continuity and it is therefore associated with more frequent fires (Camill *et al.*, 2009) and forested bogs may be more susceptible to burn as water tables are generally low. Although burning in wet, open peatlands can not be excluded under drought conditions (Cyr *et al.*, 2005), open peatlands may often remain unaffected by fire, especially when local water tables are high and trees are sparse (Hellberg, Niklasson and Granström, 2004). As a result, fire frequencies in open peatlands are generally lower than those of adjacent forest stands (Camill *et al.*, 2009 ; Kuhry, 1994 ; Zoltai *et al.*, 1998).

Peatland fires influence local C dynamics by a release of C to the atmosphere through combustion, estimated at 2.5-3.2 kg m^{-2} (Pitkänen, Turunen and Tolonen, 1999 ; Turetsky *et al.*, 2002). In addition, postfire C loss is generally important due to a delay in vegetation reestablishment (Wieder *et al.*, 2009). The negative effect of recurrent, severe fires on long-term peat and C accumulation has been established for different peatland types and regions around the globe, yet research has been more concentrated in boreal western and central Canada (Camill *et al.*, 2009 ; Kuhry, 1994 ; Robinson and Moore, 2000 ; Turetsky *et al.*, 2002). In these regions, climate-driven increases in fire frequency or severity may force peatlands to switch from net sinks to sources during the following decades (Turetsky *et al.* 2002; Wieder *et al.* 2009). Whereas these forested peatlands have developed under a dry continental climatic regime, dominance of open bogs with relatively high water tables in climatically more humid eastern boreal Canada may inhibit differential fire and C accumulation dynamics (Payette *et al.*, 1989 ; Payette and Rochefort, 2001). In the Eastmain region, peatlands have accumulated considerable amounts of C during the last ~7000 years, regionally averaging 91 kg m^{-2} at a mean rate of 16.2 g m^{-2} yr^{-1} (van Bellen *et al.*, 2011).

Here, we present a study on the Holocene patterns of fire and C sequestration based on the quantification of stratigraphic charcoal from three ombrotrophic peatlands of the Eastmain region in boreal Quebec, northeastern Canada. The main objective is to estimate if peat fires have driven long-term variations in C accumulation. Besides, we aimed to determine if peatland vegetation composition has affected peat fire regimes during the Holocene. Considering the influence of vegetation on fire regimes, we hypothesize that a higher presence of trees and shrubs is conducive to fire propagation and combustion, and thus shows a positive relationship with fire frequency and charcoal production. During the late-Holocene, peatlands in the Eastmain region were subjected to important hydrological fluctuations (van Bellen, Garneau and Booth, in press), which may have promoted fire potential. We therefore hypothesize that peatland fire frequency and charcoal production have been higher during the late-Holocene, possibly concomitant with changes in local vegetation, potentially showing a negative influence on C sequestration.

3.2. Study region

Three pristine peatlands Lac Le Caron (LLC), Mosaik (MOS) and Sterne (STE), located in the Eastmain river watershed (51°50'-52°20'N/75°00'-76°00'W) were studied (Fig. 3.1). Regional mean annual temperature is -2.1 ± 0.2°C (January: -22.0 ± 0.5°C; July: 14.6 ± 0.2°C) and mean precipitation is 735 ± 12 mm, of which about one third falls as snow (interpolated means and standard errors of 1971-2003 NLWIS data; Hutchinson *et al.*, 2009). Forest fires are important in this region with actual fire cycles (i.e. the time required to burn an area equivalent to the study area) of 90-100 years (Mansuy *et al.*, 2010 ; Payette *et al.*, 1989). Regionally, all large fires may be assumed of natural origin as human activity is highly restricted. Figure 3.2 shows an example of an Eastmain peatland burning pattern. A complete description of peatland characteristics can be found in van Bellen *et al.* (2011).

3.3. Methods

3.3.1. The use of charcoal records to reconstruct peatland fires

Long-term reconstructions of fire regimes from accumulating ecosystems as lakes and peatlands are predominantly based on macroscopic charcoal quantification from sediments, with peaks in the number of charcoal fragments possibly reflecting regional fires (Ali *et al.*, 2009). Stratigraphic charcoal presence is generally quantified by charcoal accumulation rates (CHAR; expressed as pieces cm^{-2} yr^{-1}).

Here, we assumed that all fires recorded have burned upland forest and that some of those fires may have eventually spread into the peatland. As we aimed to verify past potential effects of peatland burning on C sequestration, we then needed to convert the obtained record of forest fire events to a record of peatland burning. Determining the distance of the charcoal source is complicated by the fact that dispersal and deposition mechanisms are spatially and temporally variable depending on fire intensity, forest type, wind speed, and local topography. In general, small charcoal particles tend to travel farther than large particles (Clark, 1988 ; Lynch, Clark and Stocks, 2004 ; Ohlson and Tryterud, 2000); therefore,

quantifying large fragment presence could be a means of establishing a local fire history. However, estimates of charcoal dispersal potential are highly variable. From experimental burning in boreal Scandinavia, Ohlson and Tryterud (2000) state that practically no particles >0.5 mm are deposited outside the burning area, although in other settings 1-cm charcoal fragments have been found to be transported over distances of several kilometers (Tinner *et al.*, 2006). Due to the lack of a uniform relation between the size of a charcoal fragment and its deposition distance, a distinct size threshold, separating 'local' from 'regional' origins, can not be determined. An additional complication lies in the fact that peatland burning leaves patchy patterns (Fig. 3.2), hence burning variability within a peatland is highly important.

In this study, fire history was reconstructed at two levels. First, we identified series of charcoal peaks representing individual fires that burned at an estimated distance of 500-1000 m of the coring site (Higuera *et al.*, 2007). As coring was performed in proximity to the upland forest, these fires are assumed to generally represent the upland forest fire regime (hereafter referred to as F-fires). Second, each series was partitioned by the median charcoal peak amplitude for each peatland in order to distinguish the largest peaks. Charcoal peaks may be a proxy for fire size, fuel consumption (Higuera et al., 2009) and fire proximity and intensity (Hély *et al.*, 2010 ; Higuera, Whitlock and Gage, 2010 ; Whitlock *et al.*, 2006). Interpreting peak magnitude as an indication of fire proximity, we hypothesized that the larger peaks correspond to fires that continued from the upland forest into the peatland (FP-fires). The FP-fire record was compared to temporal trends in C accumulation rates. Given the evident negative influence of individual fire events on the C balance (Wieder *et al.*, 2009), we assumed that a significant influence of fire regimes on long-term C sequestration would be manifested by a negative relation between fire and C accumulation rates, as shown by Kuhry (1994). We defined fire frequency by the number of recorded fires per unit time, while the mean fire interval represents the average of all fire intervals from all cores (Payette, 1992).

3.3.2. Fieldwork

From each of the three peatlands two coring sites were selected. Coring locations were selected at opposing sides within each peatland (Fig. 3.1). To obtain records with a sufficiently high temporal resolution, we aimed to extract cores of at least 1.5 m in length, nearby the forest-peatland boundary that was identified by absence/presence of a surface *Sphagnum* cover, and an organic horizon thickness of >40 cm (Commission canadienne de pédologie, 1998). As the slope of the peatland basin was highly variable between sites, the distance between coring location and peatland-forest limit varied between 12 and 132 m. Coring was performed using a Box corer (10×10 cm width) to sample the upper 1 m and Russian peat samplers (4.5- to 7.5-cm diameter) for deeper horizons. Monoliths were wrapped in plastic, transferred to polyvinyl chloride tubes and stored at 4°C until analysis.

3.3.3. Laboratory work

Prior to specific treatment, cores were sliced into contiguous 1 cm subsamples in the laboratory. From each slice, 2 cm^3 was retained for macrocharcoal analysis, assumed large enough to provide replicable data (Carcaillet *et al.*, 2001b). The subsample was soaked for at least 14 hours in 5% KOH and carefully rinsed through a 355-µm sieve. Material was transferred to a dish for dry analysis. Macrocharcoal fragments were counted using a binocular microscope (×25).

Peat C contents were calculated from bulk density and loss-on-ignition (LOI) analyses. The amount of organic matter (OM) was defined as the product of bulk density and LOI (Dean, 1974). Bulk density was determined from contiguous 1 cm^3 subsamples after drying for 16 hours at 105°C. Subsequently, LOI analysis was performed at 550°C for 3.5 hours (Heiri, Lotter and Lemcke, 2001). The resulting OM was converted to organic C assuming a mean of 0.5 g C g^{-1} OM (Turunen *et al.*, 2002).

Past vegetation assemblages were quantified with a binocular microscope using a semi-quantitative technique on 2-cm^3 subsamples. Fragments of either *Sphagnum* or ligneous

origin were quantified by a volume percentage of the entire sample. *Picea* and *Larix* spp. needle fragments were quantifed at a scale of 1 to 3 (1=rare, 3=frequent).

A total of 40 samples were submitted to Keck-CCAMS facility (Irvine, USA) for ^{14}C accelerator mass spectrometry (AMS) dating. Dated samples were selected either at levels of apparent charcoal peaks or, when peaks were less conspicuous, at the boundaries of zones with abundant fragments. Radiocarbon ages were calibrated using the IntCal04 calibration curve (Reimer *et al.*, 2004) within the *Bchron* software package (Haslett and Parnell, 2008). Age-depth models were constructed assuming vertical accumulation as a continuous monotonic process applying piecewise linear interpolation. All ages were expressed as calendar years before present (BP = before AD 1950). Age-depth models were based on these calibrated ages and the age of the peatland surface when cores were extracted, established at -56 cal BP (i.e. AD 2006). C accumulation rates were calculated for different sections of each core by dividing the C mass by the period of accumulation and correcting for a uniform surface area (Clymo, Turunen and Tolonen, 1998).

3.3.4. Peat fire identification

Charcoal peaks representing fire events, were identified using *CharAnalysis* (Higuera *et al.*, 2009). As peat records generally show long-term variations in accumulation rate, records were rescaled to a constant sample age resolution to reduce biases in the ability to detect significant charcoal peaks. The sample resolution was defined by the median value of each record and the resulting record was defined as $C_{interpolated}$. Then, each charcoal record was decomposed into a record of low-frequency and a record of high-frequency fluctuations in CHAR. Low-frequency variations are assumed to reflect changes in the rate of total charcoal production, secondary charcoal transport, and sediment mixing (Clark, Royall and Chumbley, 1996 ; Higuera *et al.*, 2009), and is generally referred to as $C_{background}$. The $C_{background}$ record was calculated by a locally-weighted regression (LOWESS; robust to outliers) of which the window width varied among records (discussed below). $C_{background}$ was subtracted from $C_{interpolated}$, which resulted in the new record C_{peak}, representing high-frequency variations in CHAR. The C_{peak} record was subdivided in sections defined by the window width, and for

each section C_{peak} values were separated in two components: C_{noise}, representing variations around $C_{background}$ due to sediment mixing and sampling, and C_{fire}, fluctuations in CHAR that exceed the variability in the noise distribution. The threshold separating C_{fire} from C_{noise} was set as the 99th percentile of the noise population (cf. Higuera *et al.*, 2009). To verify the ability of each record to identify peaks associated to fires, a signal-to-noise index (SNI) was determined, which was defined by the ratio of the variance within the C_{fire} distribution and the total variance in C_{peak} for the applied window width. SNI fluctuates by definition between 0 and 1, with high values indicating a clear distinction of charcoal peaks. In addition, a measure of the threshold goodness-of-fit (GOF) was established that is equal to the p-value from a test on the fitted noise distribution and C_{noise} samples. Large p-values (>0.1) indicate that the C_{noise} distribution is significantly represented by a Gaussian distribution (Higuera *et al.*, 2009). The best smoothing window width for $C_{background}$ for each record was selected based on the optimal sum of the GOF of the noise distribution and SNI.

The median of the resulting charcoal peak values (i.e. the number of fragments associated to a peak, hereafter referred to as peak magnitude and expressed as pieces cm^{-2} $peak^{-1}$) of each peatland was used to separate FP- from F-fires. Holocene FP- and F-fire frequency was defined for each core individually as the ratio of the number of fires and the total age of peat accumulation and expressed as number of fires per 1000 years. Holocene variations in fire frequency were quantified using a 1000-year interval, after which values were smoothed with a 1500-year LOWESS filter using *KaleidaGraph 3.6*. The 1500-year filter represented the finest resolution capable of eliminating sudden shifts on all curves.

To distinguish regional tendencies, temporal variations in F- and FP-fire frequency were synthesized. Each record was standardized by subtracting the mean of that record from each single value and dividing by the record's standard deviation (i.e. z-scores).

To verify the relationship between FP-fire regimes and C accumulation, we calculated a mean frequency and fire associated-CHAR per 1000-year time slice in order to obtain a constant sample resolution for regression analyses.

3.4. Results

3.4.1. Chronologies

Radiocarbon dating showed ages of peat inception varying between 4586 (LLC_L4) and 7321 cal BP (MOS_L1) (Table 3.1). Peat accumulation rates varied during the Holocene, with generally lowest values between 2000 and 1000 cal BP, and apparent high rates towards the surface (Fig. 3.3), which is due to the undecomposed state of the subsurface litter. As core lengths were variable and correlated poorly with the age of peat inception, median sample resolutions, which were used to rescale each record, varied between 22 and 48 yrs sample^{-1} (Table 3.2).

3.4.2. Holocene fire characteristics

Identification of charcoal peaks was based on background charcoal smoothing window widths between 700 and 1000 years (Fig. 3.4), with a corresponding median SNI between 0.72 and 0.95 (Table 3.2). The maximum sum of median SNI and GOF values varied between 1.3-1.6 (Fig. 3.4).

Total number of F-fires detected varied between 10 and 19 individual events since local peat inception (Table 3.2), representing a mean (±SE) Holocene F-fire interval of 439 years (±64), or a mean frequency of 2.5 fires 1000 yrs^{-1}. Peak magnitudes ranged from 0.09 to 403.67 fragments, with generally substantially higher values in MOS_L1, MOS_L4 and STE_L4 than in LLC_L1 and LLC_L4 (Fig. 3.5; Table 3.2). The total number of FP-fires registered within each core varied between 6 and 10. A t-test on all FP-fires before and after 2000 cal BP showed that peak magnitudes after 2000 cal BP were significantly higher than those before 2000 cal BP ($p=0.03$; Table 3.2), indicating a late-Holocene trend towards more intense or severe fires. Mean FP-fire frequency varied between 2.2 (LLC_L4) and 0.8 events per 1000 years (STE_L4) (Table 3.2), representing a mean fire interval of 883 years (±105). Temporal trends in fire frequency are highly variable. LLC_L4 shows more frequent F-fire events between 4000 and 3000 cal BP, whereas only two F-fires have been registered in

LLC_L1 during this period (Fig. 3.5). In STE_L2, FP-fire frequency decreased gradually between 4000 and 1300 cal BP, while STE_L4 shows no distinct temporal variation. Compared to LLC and STE bogs, MOS bog shows higher mean peak magnitude (Table 3.2). Thus, fire regimes have been highly variable within and among ecosystems. The spatially varying trends in both F- and FP-fire frequency within each peatland imply high incidence of local-stand factors on fire occurrence and propagation.

At the regional scale, relative F and FP-fire frequency trends diverged after 2200 cal BP (Fig. 3.6a). Whereas both F- and FP-fire frequency decreased between 3400 and 2300 cal BP, FP-fires became relatively more frequent after 2200 cal BP, although F-fire frequency continued to decrease. Moreover, the subset of FP-fire peak magnitudes generally culminated around 1000 cal BP and decreased sharply after 700 cal BP (Fig. 3.6b).

3.4.3. C sequestration patterns

Holocene apparent rates of C accumulation in the lateral sections of the peatlands varied between 10.3 and 19.4 g m^{-2} yr^{-1} (Table 3.2). Temporal variations in C accumulation rates were reconstructed within all cores, and showed generally minimum C accumulation rates around 2000 cal BP between 7.3 and 12.1 g m^{-2} yr^{-1} (Fig. 3.7). Increasing apparent accumulation rates during the recentmost centuries are the result of an incomplete decomposition in the upper, temporally aerated, section of the peat.

C accumulation data, FP-CHAR and FP-fire frequency were rescaled to 1000-year time slices for regression (Fig. 3.8a-c), a period that was chosen as it is close to the mean regional Holocene FP-fire frequency of 883 years. Results show no relationship between FP-CHAR and C accumulation rate (Fig. 3.8d) and a significant but weak positive relationship between FP-fire frequency and C accumulation rate (Fig. 3.8e).

3.4.4. Vegetation reconstructions

As most important changes in both fire regimes and C accumulation have occurred towards the late-Holocene, vegetation reconstructions were focused on this period. The 3500-500 cal BP period covers both the interval with generally less frequent and low magnitude FP-fires prior to ~2000 cal BP and the interval with more frequent, high-magnitude FP-fires between 2000 and 500 cal BP. All sites show a slightly fluctuating presence of ligneous vegetation at centennial timescales, but high magnitude fires were not closely associated to important presence of ligneous vegetation (Fig. 3.9). In addition, sites with high-magnitude FP-fires during the late-Holocene had either complete absence (e.g. MOS_L1) or important presence (MOS_L4) of *Sphagnum* vegetation. Hence, the trend towards higher FP-frequency and peak magnitude between 2000 and 500 cal BP appears to have been independent of variations in local vegetation.

3.5. Discussion

3.5.1. Factors and scales in Holocene fire regimes

High median SNI values of 0.72-0.95 and high maximum sums of median SNI and GOF values of 1.3-1.6 for all records show that charcoal peaks, associated to the F-fire records, were clearly separated from background values (Higuera *et al.*, 2009). Although charcoal peaks could not be validated as no independent past regional fire records were available, the charcoal deposition pattern of a recent fire aided interpretation. In 1995, the adjacent forest along the northwestern section of MOS bog burned, as indicated by fire maps (Ministère des Ressources naturelles et de la Faune, 2010), although this fire did not affect the coring locations. Considering the apparent rapid accumulation and some potential for small fragments to move vertically through unconsolidated peat, this fire may have caused charcoal deposition in the upper 20 cm of the peat. Some concentrated charcoal fragments were found in MOS_L1 and MOS_L4 around 18 and 14 cm, respectively, yet *CharAnalysis*

identified these charcoal peaks as insignificant. Thus, these fires were correctly omitted in the reconstructed fire record.

The reconstructed mean Holocene F-fire interval of 439 years, equivalent to a mean frequency of 2.5 fires 1000 yrs^{-1}, is high compared to the actual mean regional fire cycle (i.e. the period needed to burn an area equivalent to the study area) of 90-100 years (Mansuy *et al.*, 2010 ; Payette *et al.*, 1989), implicating that our reconstructions may be rather conservative. The difference between the actual mean regional fire cycle and the mean Holocene F-fire interval from stands adjacent to peatlands may be partly due to the fact that uplands in proximity of peatlands may burn less frequently than those at higher elevation (Mansuy *et al.*, 2010). Neverthcless, somewhat underestimated peatland fire frequencies would most likely not affect the correlation with C accumulation rates: regression of F-fire frequency and C accumulation rates resulted in a weak, positive correlation ($p=0.017$, $r^2=0.16$), similar to the poor relationship between FP-fire frequency and C accumulation rates shown in Fig. 3.8e.

The mean Holocene FP-fire interval of 883 years of the Eastmain region peatlands, equivalent to a mean frequency of 1.1 fire 1000 yrs^{-1}, largely corresponds to results from other boreal regions, even though spatial and temporal variability in fire activity is relatively high. Zoltai *et al.* (1998) estimated peat fire return intervals around 150 years in continental boreal bogs and 800 years in humid boreal bogs. In Manitoba, fire return intervals were quantified between 624 and 2930 years, depending on the criterion for local fire identification (Camill *et al.*, 2009). In western Canada, reconstructions in *Sphagnum* peatlands show temporally and spatially highly varying frequencies, from mid-Holocene frequency of 5.3 fires 1000 yrs^{-1} to no recorded fires over the entire accumulation in other sections (Kuhry, 1994). In the Eastmain peatlands, charcoal horizons were visually indistinct throughout the sequences, probably because of relatively low charcoal fragment quantities and poor peat preservation resulting in a general dark appearance. The absence of charred *Sphagna* in these horizons implies that most burning was surficial, only charring the standing biomass, i.e. trees and (small) shrubs.

FP-fire regimes in the region have been influenced by multiple factors, acting on different temporal and spatial scales. Regionally uniform increasing late-Holocene fire intensity may have been forced by climatic factors (discussed below). Dissimilar FP-fire

frequencies among peatlands imply control at the landscape level, as variation in landscape connectivity (i.e. the configuration of lakes and rivers and the forest mosaic) determines both spatial and temporal patterns of the fire regime (Bergeron, 1991). Poor temporal synchroneity of FP-fire events within each peatland points towards patchy burning patterns, probably forced by spatially variable vegetation and hydrology (Fig. 3.2). These results confirm that, in order to obtain complete understanding of the complexity of peatland fire regimes, multiple records from multiple peatlands are essential to cover spatial variability.

Gavin *et al.* (2003) suggest that an apparent regional increase in peak magnitude during the late-Holocene might be explained by higher decomposition of charcoal fragments towards older deposits. This is probably of minor importance in our study as peak fragments were generally large (>0.5 mm) and decomposed organic matter did not show a distinct trend downward in detailed stratigraphic analyses (van Bellen, Garneau and Booth, in press).

Eastmain peatlands are characterized by important microtopography of high hummocks (~40 cm height) with frequent *Picea mariana* presence and hollows with standing water, resulting in spatially variable vegetation assemblages and hydrology. As centennial-scale shifts in vegetation succession in the stratigraphies did not show any clear link with individual fire events, we hypothesize that peat fires have been the result of shifts in water tables independent of important changes in vegetation. Multiple climate-forced episodes of low water tables have been reconstructed in the region since 3000 cal BP (Fig. 3.9; van Bellen, Garneau and Booth, in press). These periods were identified by stratigraphic analyses of testate amoeba assemblages at a decadal- to centennial-scale resolution. Some of the FP-fires might correlate with the regional episodes of dry conditions (Fig. 3.9), although the generally large 95%-confidence interval of the chronologies obscures a potential link (Fig. 3.3). High-resolution dating strategies should be adopted for confirmation of eventual peatland fire sensitivity to decadal- to centennial-scale dry shifts. Alternatively, only a couple of years of dry climate conditions may be sufficient for a peatland to burn and therefore correlation with decadal- to centennial-scale droughts may be poor, even if high-resolution dating is applied. Additional complexity is caused by the low median sample resolution (22-48 yrs sample^{-1}), implying that such short-term changes would be hardly detectable by stratigraphic analyses.

3.5.2. Fire regimes and C sequestration

The absence of a negative correlation between both FP-fire frequency and FP-CHAR, and C accumulation rates, indicates that Holocene C sequestration has not been primarily driven by fire regime. As fire events have by definition a negative effect on C sequestration through direct emissions and postfire net emissions (Turetsky *et al.*, 2002 ; Wieder *et al.*, 2009), fire effects have apparently been masked by other, more important autogenic or allogenic factors. General low FP-fire frequencies in the Eastmain region, combined to surficial fire types, may be the principal cause for the absence of a significant influence of fire regimes on long-term C accumulation rates.

C accumulation patterns are determined by the difference between primary production and decomposition, which generally depend on climate regime (e.g. temperature, precipitation and moisture balance effects) and internal peatland dynamics (large-scale peatland surface topography, microtopography and basin morphology) (Belyea and Clymo, 2001 ; Belyea and Malmer, 2004 ; van Bellen *et al.*, 2011 ; Yu, Beilman and Jones, 2009). Eastmain peatland long-term C accumulation has been influenced by late-Holocene climatic fluctuations (van Bellen, Garneau and Booth, in press) that were mediated by local geomorphology (basin morphology) and surface topography factors (van Bellen *et al.*, 2011). Climatic cooling may have been the principal driving factor of C sequestration, as limited rates have been identified between 2000 and 1000 cal BP in multiple cores from the Eastmain region (van Bellen *et al.*, 2011 ; van Bellen, Garneau and Booth, in press). As fires generally originate from uplands and spread into peatlands, burning should be more frequent near the forest-peatland boundary compared to the central parts (e.g. as shown in Fig. 3.2). Although lateral coring positions may thus be advantageous for fire reconstructions, these sections are possibly also more affected by autogenic change. For instance, the forest-peatland boundary, forming an ecotone of varying width, is probably more frequently exposed to (episodic) minerotrophic input, paludification and concomitant changes in vegetation and it may have a lower vegetation resilience (Bauer *et al.*, 2009 ; Bhatti *et al.*, 2006) than central sections of the peatland. Thus, fire effects on C accumulation may well have been minimized by a more important local effect of autogenic factors.

3.5.3. Late-Holocene climate and fire regime

Paleoecological reconstructions from the northeastern Canadian boreal forest and forest-tundra show a general cooling trend with more frequent forest fires from 3000 cal BP (Asselin and Payette, 2005 ; Payette and Gagnon, 1985), culminating between 2000 and 1000 cal BP (Arseneault and Sirois, 2004 ; Asselin and Payette, 2005 ; Carcaillet *et al.*, 2010). In the southern boreal forest, some contrasting trends with either lower or higher fire frequency have been found, possibly resulting from regional variations in the sensitivity of the fire regime to climatic variations during the late-Holocene (Ali *et al.*, 2008 ; Ali, Carcaillet and Bergeron, 2009 ; Carcaillet *et al.*, 2001a ; Hély *et al.*, 2010). In this study, diverging trends in F- and FP-fire frequency after 2200 cal BP might imply a differential forcing. Alternatively, a climatic influence on regional fire frequency in both uplands and peatlands may be explained by differential ecosystem response. In forest stands, climate primarily regulates large-scale fire activity by the occurrence of summer drought events with variations in temperature, precipitation and wind speed (Bergeron and Archambault, 1993 ; Senici *et al.*, in press). However, in open peatlands, drainage and insolation dynamics, and thus summer water table fluctuations, are linked to climate in a different manner. Summer precipitation and to a lesser extent temperature are likely to influence summer water tables (Booth, 2010 ; Charman *et al.*, 2009), but winter precipitation and temperature may play an important role in specific settings (e.g. Lamentowicz *et al.*, 2010). Furthermore, indirect climate effects and climate-vegetation interaction may exert an additional influence on peatland surface drought. Snow cover, for instance, varies with vegetation type and microtopography (Camill and Clark, 2000), and the combination of low temperatures and thin snow covers may allow permafrost aggradation (Vitt, Halsey and Zoltai, 1994), which further complicates the climate-hydrology relationship. Multiple replicate episodes of low water tables have been reconstructed in the Eastmain peatlands between 3000-100 cal BP, which were related to episodic dry and cold climatic conditions, possibly exacerbated by decadally persisting frozen peat horizons (van Bellen, Garneau and Booth, in press).

Thus, even if regional climate has been a principal factor driving long-term variations in water tables and moisture contents, upland forests and peatlands potentially show differential trends as they are driven by specific aspects of climate regimes, that are mediated

by specific ecological feedback on climate conditions (van der Molen and Wijmstra, 1994 ; Wotton and Beverly, 2007). As a result, differing trends in fire frequency between upland forests and open peatlands are probable even if climate is the sole driving factor, complicating comparisons between historical fire regimes of these two ecosystems.

3.5.4. Future perspectives on Eastmain peatland fire potential

Holocene peatland fire regimes have not been a determinant factor in long-term C sequestration in the Eastmain region, and this may well be valid for most open peatlands in boreal eastern Canada. Nevertheless, climate-fire-carbon cycling interactions may be variable in the light of present and ongoing climate change (Bergeron *et al.*, 2010). Peatland fire regimes are partly linked to adjacent upland forest fire activity and possibly short-term drought and therefore these concepts should be considered in the projection of future peatland fire regimes. Boreal Quebec climate projections, as modelled by the Canadian Regional Climate Model (A2 scenario), show increases in summer temperature of 2.0-2.5°C and ~10% increases in summer precipitation around 2050 relative to 1980 in boreal Quebec (Plummer *et al.*, 2006). As a result, fire regime scenarios for the Waswanipi region (~200 km south of the Eastmain region) indicate an increase in annual area burned of 7% and an increase in monthly fire risk of 30%, attaining 70% in July and 100% in August by 2100 (Le Goff, Flannigan and Bergeron, 2009). These increases in upland fire activity suggest a higher potential for peatland fire (Flannigan *et al.*, 2009), yet climate may have a different influence on burning potential in large, wet and open peatlands. A trend of higher water tables and reduced drought frequency since the end of the last Little Ice Age episode (~AD 1850) in both peatlands and forest stands has been observed in northern Quebec (Arlen-Pouliot and Bhiry, 2005 ; Bergeron and Archambault, 1993 ; Lesieur, Gauthier and Bergeron, 2002 ; Loisel and Garneau, 2010 ; Payette and Delwaide, 2004 ; Payette *et al.*, 2004 ; van Bellen, Garneau and Booth, in press). Following these results, climate projections for peatland fire dynamics remain uncertain in northeastern Canada. This contrasts with scenarios for western continental Canada, where peatlands presently persist at the dry climatic end of their global

distribution. A projected major increase in fire activity may cause western Canadian peatlands to switch from net sinks to sources of C (Wieder *et al.*, 2009).

3.6. Conclusion

Holocene reconstructions show that fire regimes have not been sufficiently important to have driven variations in long-term C sequestration in boreal peatlands of northeastern Canada. However, FP-fires were associated with increasing charcoal production and showed slightly higher frequencies between 2000 and 500 cal BP, a period that was also characterized by generally low rates of C accumulation of 7.3-12.1 g m^{-2} yr^{-1}.

Peatland fire regimes reconstructed in this study have generally been a function of regional (climatic), extra-local (e.g. landscape connectivity) and local (e.g. peatland microtopography) factors. As vegetation reconstructions failed to show a clear link with fire regimes, colder climate regimes and concomitant unstable hydrological conditions may have been principal factors limiting peatland C sequestration in the Eastmain region during the same period (van Bellen, Garneau and Booth, in press). A direct link between fire events and climate-driven episodic dry shifts might be established if higher-resolution dating strategies were applied to the charcoal record. Diverging trends in F-fires and FP-fires after 2200 cal BP may be explained by a differential sensitivity of local hydrology to climatic forcing. For a complete comprehension on driving factors of long-term fire regimes, the use of multiple records from multiple peatlands appears essential.

Although peatland hydrology depends on the delicate balance between temperature, precipitation and vegetation effects, 21st century warmer and wetter climatic conditions might induce rising peatland water tables and peatland fire regimes might thus remain of minor importance, although the effect of increasing upland fire frequency should be taken into account. A comparable trend is visible since the end of the Little Ice Age (~AD 1850), as peatlands have become progressively wetter and possibly less vulnerable to burning.

Acknowledgements

We thank H. Asnong, G. St-Pierre and B. Talbot for field assistance and M. Boivin, A. Tadros and A. Thibault for assistance in the laboratory. Thanks to the hydroclimatological scenarios team of the Ouranos consortium for providing climate data by manipulation of 1971-2003 NLWIS data and to P.-L. Dallaire for help with cartography. Suggestions on analyses and revision by H. Asselin greatly improved this paper. Funding was provided by Hydro-Quebec Production through the EM-1 Project *Reservoirs' net greenhouse gas emissions research project (2005-2009).*

References

Ali, A. A., H. Asselin, A. C. Larouche, Y. Bergeron, C. Carcaillet and P. J. H. Richard. 2008. «Changes in fire regime explain the Holocene rise and fall of Abies balsamea in the coniferous forests of western Quebec, Canada». *The Holocene*. vol. 18, no 5, p. 693-703.

Ali, A. A., C. Carcaillet and Y. Bergeron. 2009. «Long-term fire frequency variability in the eastern Canadian boreal forest: the influences of climate vs. local factors». *Global Change Biology*. vol. 15, no 5, p. 1230-1241.

Ali, A. A., P. E. Higuera, Y. Bergeron and C. Carcaillet. 2009. «Comparing fire-history interpretations based on area, number and estimated volume of macroscopic charcoal in lake sediments». *Quaternary Research*. vol. 72, no 3, p. 462-468.

Arlen-Pouliot, Y., and N. Bhiry. 2005. «Palaeoecology of a palsa and a filled thermokarst pond in a permafrost peatland, subarctic Québec, Canada». *The Holocene*. vol. 15, no 3, p. 408-419.

Arseneault, D., and L. Sirois. 2004. «The millennial dynamics of a boreal forest stand from buried trees». *Journal of Ecology*. vol. 92, p. 490-504.

Asselin, H., and S. Payette. 2005. «Late Holocene opening of the forest tundra landscape in northern Québec, Canada». *Global Ecology and Biogeography*. vol. 14, no 4, p. 307-313.

Bauer, I. E., J. S. Bhatti, C. Swanston, R. K. Wieder and C. M. Preston. 2009. «Organic matter accumulation and community change at the peatland-upland interface: Inferences from ^{14}C and ^{210}Pb dated profiles». *Ecosystems*. vol. 12, no 4, p. 636-653.

Belyea, L. R., and R. S. Clymo. 2001. «Feedback control of the rate of peat formation». *Proceedings: Biological Sciences*. vol. 268, no 1473, p. 1315-1321.

Belyea, L. R., and N. Malmer. 2004. «Carbon sequestration in peatland: patterns and mechanisms of response to climate change». *Global Change Biology*. vol. 10, no 7, p. 1043-1052.

Bergeron, Y. 1991. «The influence of island and mainland lakeshore landscapes on boreal forest fire regimes». *Ecology*. vol. 72, no 6, p. 1980-1992.

Bergeron, Y., and S. Archambault. 1993. «Decreasing frequency of forest fires in the southern boreal zone of Quebec and its relation to global warming since the end of the 'Little Ice Age'». *The Holocene*. vol. 3, no 3, p. 255-259.

Bergeron, Y., D. Cyr, M. P. Girardin and C. Carcaillet. 2010. «Will climate change drive 21st century burn rates in Canadian boreal forests outside of natural variability: collating global climate model experiments with sedimentary charcoal data». *International Journal of Wildland Fire*.

Bhatti, J. S., R. C. Errington, I. E. Bauer and P. A. Hurdle. 2006. «Carbon stock trends along forested peatland margins in central Saskatchewan». *Canadian Journal of Soil Science*. vol. 86, no 2, p. 321-333.

Booth, R. K. 2010. «Testing the climate sensitivity of peat-based paleoclimate reconstructions in mid-continental North America». *Quaternary Science Reviews*. vol. 29, no 5-6, p. 720-731.

Camill, P., A. Barry, E. Williams, C. Andreassi, J. Limmer and D. Solick. 2009. «Climate-vegetation-fire interactions and their impact on long-term carbon dynamics in a boreal peatland landscape in northern Manitoba, Canada». *Journal of Geophysical Research*. vol. 114.

Camill, P., and J. S. Clark. 2000. «Long-term perspectives on lagged ecosystem responses to climate change: permafrost in boreal peatlands and the grassland/woodland boundary». *Ecosystems*. vol. 3, no 6, p. 534-544.

Carcaillet, C., Y. Bergeron, P. J. H. Richard, B. Fréchette, S. Gauthier and Y. T. Prairie. 2001a. «Change of fire frequency in the eastern Canadian boreal forests during the Holocene: does vegetation composition or climate trigger the fire regime?». *Journal of Ecology*. vol. 89, no 6, p. 930-946.

Carcaillet, C., M. Bouvier, B. Fréchette, A. C. Larouche and P. J. H. Richard. 2001b. «Comparison of pollen-slide and sieving methods in lacustrine charcoal analyses for local and regional fire history». *The Holocene*. vol. 11, no 4, p. 467-476.

Carcaillet, C., P. J. H. Richard, Y. Bergeron, B. Fréchette and A. A. Ali. 2010. «Resilience of the boreal forest in response to Holocene fire-frequency changes assessed by pollen diversity and population dynamics». *International Journal of Wildland Fire*. vol. 19, p. 1026-1039.

Charman, D. J., K. E. Barber, M. Blaauw, P. G. Langdon, D. Mauquoy, T. J. Daley, P. D. M. Hughes and E. Karofeld. 2009. «Climate drivers for peatland palaeoclimate records». *Quaternary Science Reviews*. vol. 28, no 19-20, p. 1811-1819.

Clark, J. S. 1988. «Particle motion and the theory of charcoal analysis: Source area, transport, deposition, and sampling». *Quaternary Research*. vol. 30, no 1, p. 67-80.

Clark, J. S., P. D. Royall and C. Chumbley. 1996. «The role of fire during climate change in an eastern deciduous forest at Devil's Bathtub, New York». *Ecology*. vol. 77, no 7, p. 2148-2166.

Clymo, R. S., J. Turunen and K. Tolonen. 1998. «Carbon accumulation in peatland». *Oikos*. vol. 81, no 2, p. 368-388.

Commission canadienne de pédologie. (1998). «Le système canadien de classification des sols, 3è édition». Direction de la Recherche. Ottawa, Ministère de l'Agriculture du Canada. Publication 1646: 187 p.

Conny, J. M., and J. F. Slater. 2002. «Black carbon and organic carbon in aerosol particles from crown fires in the Canadian boreal forest». *Journal of Geophysical Research.* vol. 107, no D11, p. 4116.

Cyr, D., Y. Bergeron, S. Gauthier and A. C. Larouche. 2005. «Are the old-growth forests of the Clay Belt part of a fire-regulated mosaic?». *Canadian Journal of Forest Research.* vol. 35, no 1, p. 65-73.

Dean, W. E. 1974. «Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition; comparison with other methods». *Journal of Sedimentary Research.* vol. 44, no 1, p. 242-248.

Flannigan, M., B. Stocks, M. Turetsky and M. Wotton. 2009. «Impacts of climate change on fire activity and fire management in the circumboreal forest». *Global Change Biology.* vol. 15, no 3, p. 549-560.

Gavin, D. G., L. B. Brubaker and K. P. Lertzman. 2003. «Holocene fire history of a coastal temperate rain forest based on soil charcoal radiocarbon dates». *Ecology.* vol. 84, no 1, p. 186-201.

Haslett, J., and A. Parnell. 2008. «A simple monotone process with application to radiocarbon-dated depth chronologies». *Journal of the Royal Statistical Society: Series C.* vol. 57, no 4, p. 399-418.

Heiri, O., A. F. Lotter and G. Lemcke. 2001. «Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results». *Journal of Paleolimnology.* vol. 25, no 1, p. 101-110.

Hellberg, E., M. Niklasson and A. Granström. 2004. «Influence of landscape structure on patterns of forest fires in boreal forest landscapes in Sweden». *Canadian Journal of Forest Research.* vol. 34, p. 332-338.

Hély, C., M. P. Girardin, A. A. Ali, C. Carcaillet, S. Brewer and Y. Bergeron. 2010. «Eastern boreal North American wildfire risk of the past 7000 years: A model-data comparison». *Geophysical Research Letters.* vol. 37, no 14, p. L14709.

Higuera, P. E., L. B. Brubaker, P. M. Anderson, F. S. Hu and T. A. Brown. 2009. «Vegetation mediated the impacts of postglacial climate change on fire regimes in the south-central Brooks Range, Alaska». *Ecological Monographs.* vol. 79, no 2, p. 201-219.

Higuera, P. E., M. E. Peters, L. B. Brubaker and D. G. Gavin. 2007. «Understanding the origin and analysis of sediment-charcoal records with a simulation model». *Quaternary Science Reviews.* vol. 26, no 13-14, p. 1790-1809.

Higuera, P. E., C. Whitlock and J. A. Gage. 2010. «Linking tree-ring and sediment-charcoal records to reconstruct fire occurrence and area burned in subalpine forests of Yellowstone National Park, USA». *The Holocene.*

Hutchinson, M. F., D. W. McKenney, K. Lawrence, J. H. Pedlar, R. F. Hopkinson, E. Milewska and P. Papadopol. 2009. «Development and testing of Canada-wide interpolated spatial models of daily minimum–maximum temperature and precipitation for 1961–2003». *Journal of Applied Meteorology and Climatology.* vol. 48, p. 725-741.

Korhola, A. 1994. «Radiocarbon evidence for rates of lateral expansion in raised mires in southern Finland». *Quaternary Research.* vol. 42, no 3, p. 299-307.

Korhola, A., M. Ruppel, H. Seppä, M. Väliranta, T. Virtanen and J. Weckström. 2010. «The importance of northern peatland expansion to the late-Holocene rise of atmospheric methane». *Quaternary Science Reviews*. vol. 29, no 5-6, p. 611-617.

Kuhry, P. 1994. «The role of fire in the development of Sphagnum-dominated peatlands in western boreal Canada». *Journal of Ecology*. vol. 82, no 4, p. 899-910.

Lamentowicz, M., W. van der Knaap, L. Lamentowicz, J. F. N. van Leeuwen, E. A. D. Mitchell, T. Goslar and C. Kamenik. 2010. «A near-annual palaeohydrological study based on testate amoebae from a sub-alpine mire: Surface wetness and the role of climate during the instrumental period». *Journal of Quaternary Science*. vol. 25, no 2, p. 190-202.

Le Goff, H., M. D. Flannigan and Y. Bergeron. 2009. «Potential changes in monthly fire risk in the eastern Canadian boreal forest under future climate change». *Canadian Journal of Forest Research*. vol. 39, p. 2369-2380.

Lesieur, D., S. Gauthier and Y. Bergeron. 2002. «Fire frequency and vegetation dynamics for the south-central boreal forest of Quebec, Canada». *Canadian Journal of Forest Research*. vol. 32, no 11, p. 1996-2009.

Loisel, J., and M. Garneau. 2010. «Late Holocene paleoecohydrology and carbon accumulation estimates from two boreal peat bogs in eastern Canada: Potential and limits of multi-proxy archives». *Palaeogeography, Palaeoclimatology, Palaeoecology*. vol. 291, no 3-4, p. 493-533.

Lynch, J. A., J. S. Clark and B. J. Stocks. 2004. «Charcoal production, dispersal, and deposition from the Fort Providence experimental fire: Interpreting fire regimes from charcoal records in boreal forests». *Canadian Journal of Forest Research*. vol. 34, no 8, p. 1642-1656.

MacDonald, G. M., D. W. Beilman, K. V. Kremenetski, Y. Sheng, L. C. Smith and A. A. Velichko. 2006. «Rapid early development of circumarctic peatlands and atmospheric CH_4 and CO_2 variations». *Science*. vol. 314, no 5797, p. 285-288.

Mansuy, N., S. Gauthier, A. Robitaille and Y. Bergeron. 2010. «The effects of surficial deposit-drainage combinations on spatial variations of fire cycles in the boreal forest of eastern Canada.». *International Journal of Wildland Fire*. vol. 19, no 8, p. 1083-1098.

Ministère des Ressources naturelles et de la Faune. (2010). «Données historiques sur les feux de forêt au Québec». Gouvernement du Québec, Direction de l'environnement et de la protection des forêts.

Ohlson, M., and E. Tryterud. 2000. «Interpretation of the charcoal record in forest soils: forest fires and their production and deposition of macroscopic charcoal». *The Holocene*. vol. 10, no 4, p. 519-525.

Payette, S. 1992. «Fire as a controlling process in the North American boreal forest». In *A systems analysis of the global boreal forest,* H. H. Shugart, R. Leemans et G. B. Bonan, p. 144-169. Cambridge: Cambridge University Press.

Payette, S., and A. Delwaide. 2004. «Dynamics of subarctic wetland forests over the past 1500 years ». *Ecological Monographs*. vol. 74, no 3, p. 373-391.

Payette, S., A. Delwaide, M. Caccianiga and M. Beauchemin. 2004. «Accelerated thawing of subarctic peatland permafrost over the last 50 years». *Geophysical Research Letters*. vol. 31, no 18, p. L18208.

Payette, S., and R. Gagnon. 1985. «Late Holocene deforestation and tree regeneration in the forest-tundra of Quebec». *Nature*. vol. 313, no 6003, p. 570-572.

Payette, S., C. Morneau, L. Sirois and M. Desponts. 1989. «Recent fire history of the northern Quebec biomes». *Ecology*. vol. 70, no 3, p. 656-673.

Payette, S., and L. Rochefort. 2001. *Écologie des tourbières du Québec-Labrador* Trad. de: *French*. Ste-Foy: Les presses de l'Université Laval, 621 p.

Pitkänen, A., J. Turunen and K. Tolonen. 1999. «The role of fire in the carbon dynamics of a mire, eastern Finland». *The Holocene*. vol. 9, no 4, p. 453-462.

Plummer, D. A., D. Caya, A. Frigon, H. Côté, M. Giguère, D. Paquin, S. Biner, R. Harvey and R. de Elia. 2006. «Climate and climate change over North America as simulated by the Canadian RCM». *Journal of Climate*. vol. 19, no 13, p. 3112-3132.

Reimer, P. J., M. G. L. Baillie, E. Bard, A. Bayliss, J. W. Beck, C. J. H. Bertrand, P. G. Blackwell, C. E. Buck, G. S. Burr, K. B. Cutler, P. E. Damon, R. L. Edwards, R. G. Fairbanks, M. Friedrich, T. P. Guilderson, A. G. Hogg, K. A. Hughen, B. Kromer, G. McCormac, S. Manning, C. Bronk Ramsey, R. W. Reimer, S. Remmele, J. R. Southon, M. Stuiver, S. Talamo, F. W. Taylor, J. van der Plicht and C. E. Weyhenmeyer. 2004. «INTCAL04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP». *Radiocarbon*. vol. 46, no 3, p. 1029-1058.

Robinson, S. D., and T. R. Moore. 2000. «The influence of permafrost and fire upon carbon accumulation in high boreal peatlands, Northwest Territories, Canada». *Arctic, Antarctic, and Alpine Research*. vol. 32, no 2, p. 155-166.

Senici, D., H. Y. H. Chen, Y. Bergeron and D. Cyr. in press. «Spatiotemporal variations of fire frequency in central boreal forest». *Ecosystems*, p. 1-12.

Tinner, W., S. Hofstetter, F. Zeugin, M. Conedera, T. Wohlgemuth, L. Zimmermann and R. Zweifel. 2006. «Long-distance transport of macroscopic charcoal by an intensive crown fire in the Swiss Alps - implications for fire history reconstruction». *The Holocene*, p. 287-292.

Tolonen, K., and J. Turunen. 1996. «Accumulation rates of carbon in mires in Finland and implications for climate change». *The Holocene*. vol. 6, no 2, p. 171-178.

Turetsky, M., K. Wieder, L. Halsey and D. Vitt. 2002. «Current disturbance and the diminishing peatland carbon sink». *Geophysical Research Letters*. vol. 29, no 11, 1526, p. doi:10.1029/2001GL014000.

Turunen, J., E. Tomppo, K. Tolonen and A. Reinikainen. 2002. «Estimating carbon accumulation rates of undrained mires in Finland-application to boreal and subarctic regions». *The Holocene*. vol. 12, no 1, p. 69-80.

van Bellen, S., P.-L. Dallaire, M. Garneau and Y. Bergeron. 2011. «Quantifying spatial and temporal Holocene carbon accumulation in ombrotrophic peatlands of the Eastmain region, Quebec, Canada». *Global Biogeochemical Cycles*.

van Bellen, S., M. Garneau and R. K. Booth. in press. «Holocene carbon accumulation rates from three ombrotrophic peatlands in boreal Quebec, Canada: impact of climate-driven ecohydrological change». *The Holocene*.

van der Molen, P. C., and T. A. Wijmstra. 1994. «The thermal regime of hummock-hollow complexes on Clara bog, co. Offaly». *Biology and Environment: Proceedings of the Royal Irish Academy*. vol. 94B, no 3, p. 209-221.

Vitt, D. H., L. A. Halsey and S. C. Zoltai. 1994. «The bog landforms of continental Western Canada in relation to climate and permafrost patterns». *Arctic and Alpine Research*. vol. 26, no 1, p. 1-13.

Whitlock, C., M. M. Bianchi, P. J. Bartlein, V. Markgraf, J. Marlon, M. Walsh and N. McCoy. 2006. «Postglacial vegetation, climate, and fire history along the east side of the Andes (lat 41-42.5°S), Argentina». *Quaternary Research*. vol. 66, no 2, p. 187-201.

Wieder, R. K., K. D. Scott, K. Kamminga, M. A. Vile, D. H. Vitt, T. Bone, B. Xu, B. W. Benscoter and J. S. Bhatti. 2009. «Postfire carbon balance in boreal bogs of Alberta, Canada». *Global Change Biology*. vol. 15, no 1, p. 63-81.

Wotton, B. M., and J. L. Beverly. 2007. «Stand-specific litter moisture content calibrations for the Canadian Fine Fuel Moisture Code». *International Journal of Wildland Fire*. vol. 16, no 4, p. 463-472.

Yu, Z., D. W. Beilman and M. C. Joncs. 2009. «Sensitivity of northern peatland carbon dynamics to Holocene climate change». In *Carbon cycling in northern peatlands,* Andrew J. Baird, Lisa R. Belyea, Xavier Comas, A. S. Reeve et Lee D. Slater, p. 55-69. Washington: American Geophysical Union.

Yu, Z., J. Loisel, D. P. Brosseau, D. W. Beilman and S. J. Hunt. 2010. «Global peatland dynamics since the Last Glacial Maximum». *Geophysical Research Letters*. vol. 37, no 13, p. L13402.

Zoltai, S. C., L. A. Morrissey, G. P. Livingston and W. J. de Groot. 1998. «Effects of fires on carbon cycling in North American boreal peatlands». *Environmental Reviews*. vol. 6, no 1, p. 13-24.

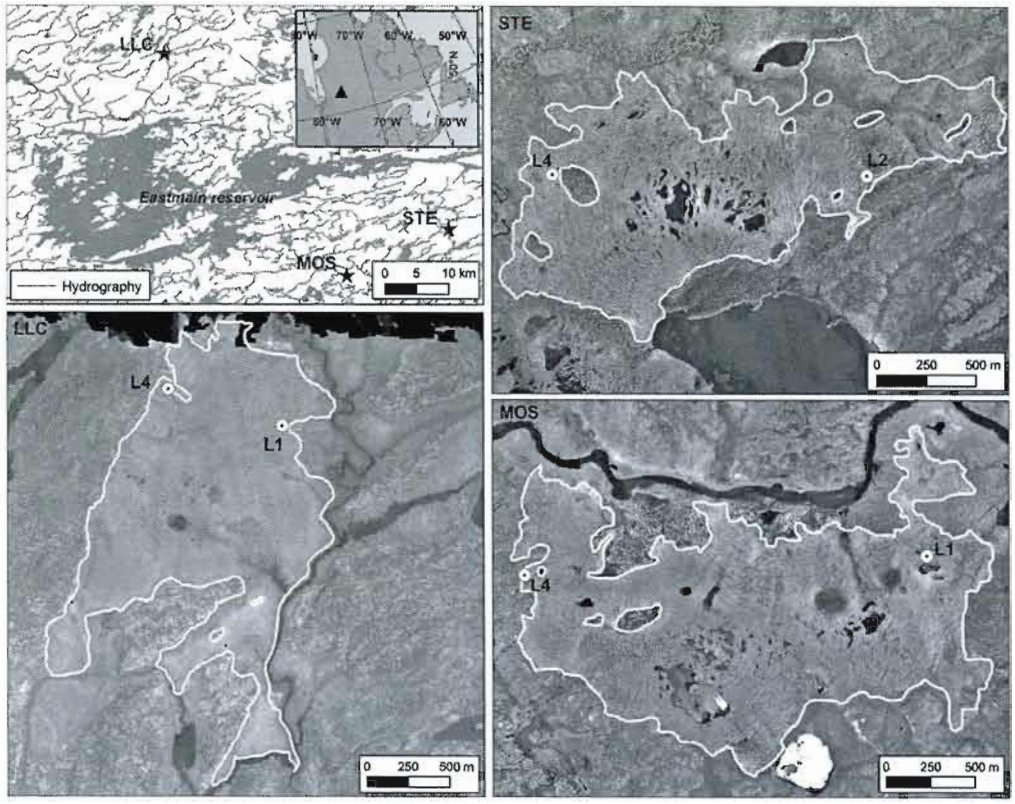

Figure 3.1. Study region, peatlands and coring locations.

Figure 3.2. Eastmain region peatland fire. This peatland was not a study site. Photo by Hydro-Québec.

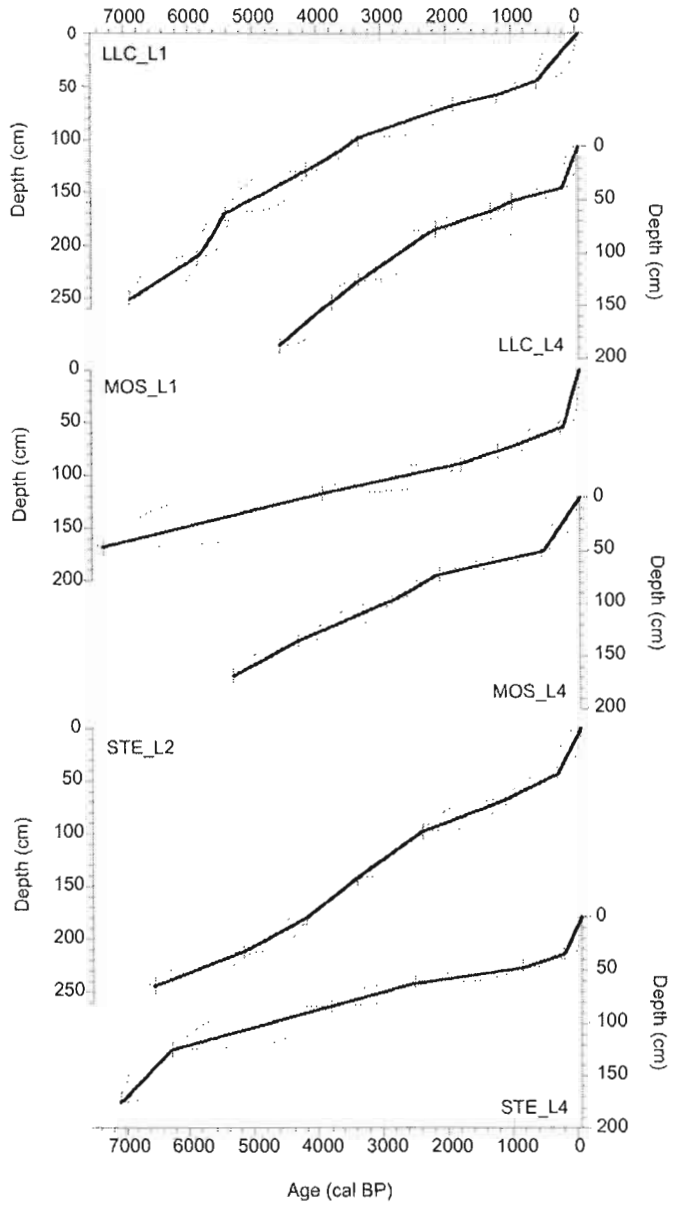

Figure 3.3. Age-depth model (black) and 95% age confidence intervals (grey, dashed) for each record.

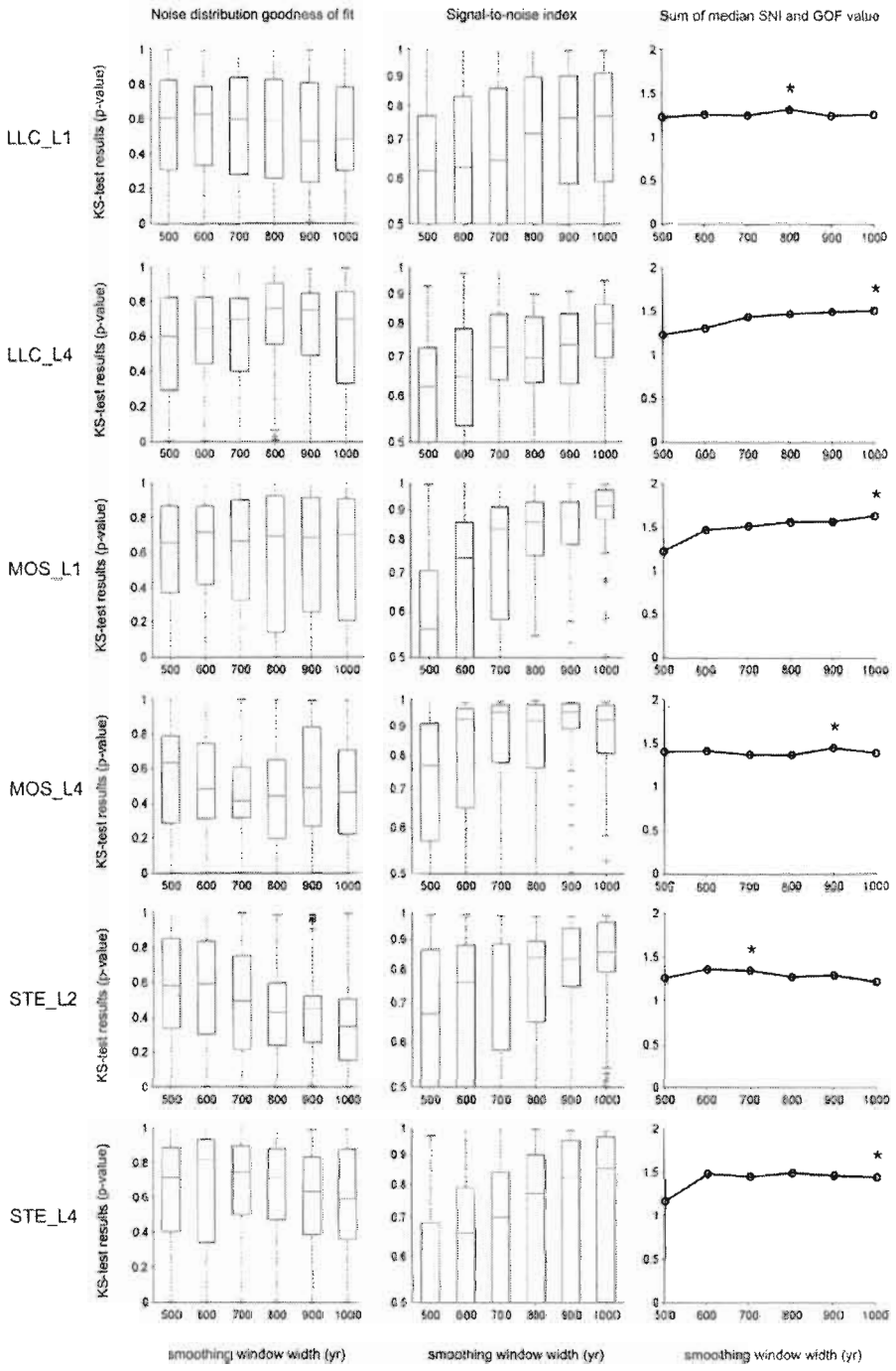

Figure 3.4. Sensitivity to different window widths for $C_{background}$ for each record. The maximum sums of median values that determined the selected smoothing window width are marked with an asterisk.

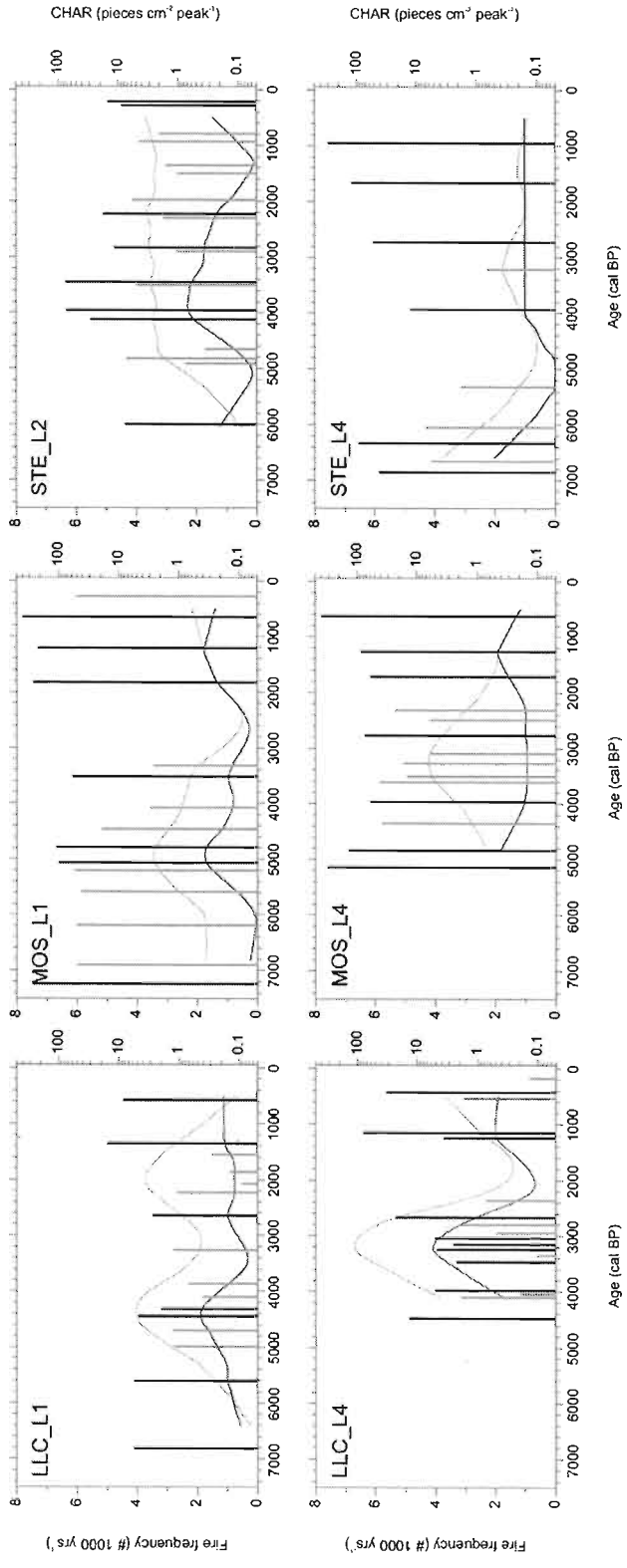

Figure 3.5. F- (grey) and FP (black)-fire frequency and peak magnitude during the Holocene for each core. Vertical bars indicate peak magnitude. Curves indicate fire frequency.

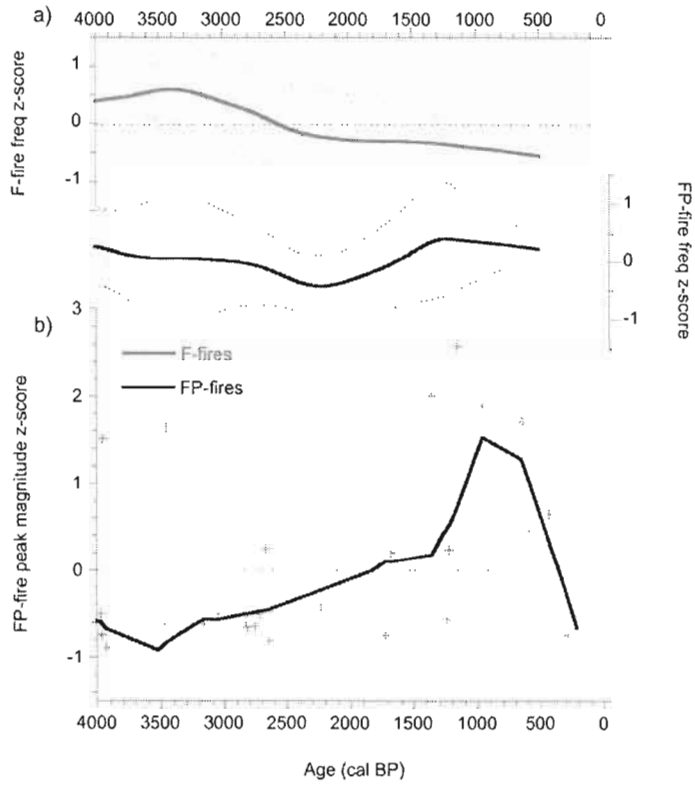

Figure 3.6. F- and FP-fire regime reconstructions since 4000 cal BP. (a) F- and FP-fire frequency. Dotted lines indicate 95% confidence interval. (b) FP-fire peak magnitudes. Solid line represents LOWESS smoothing.

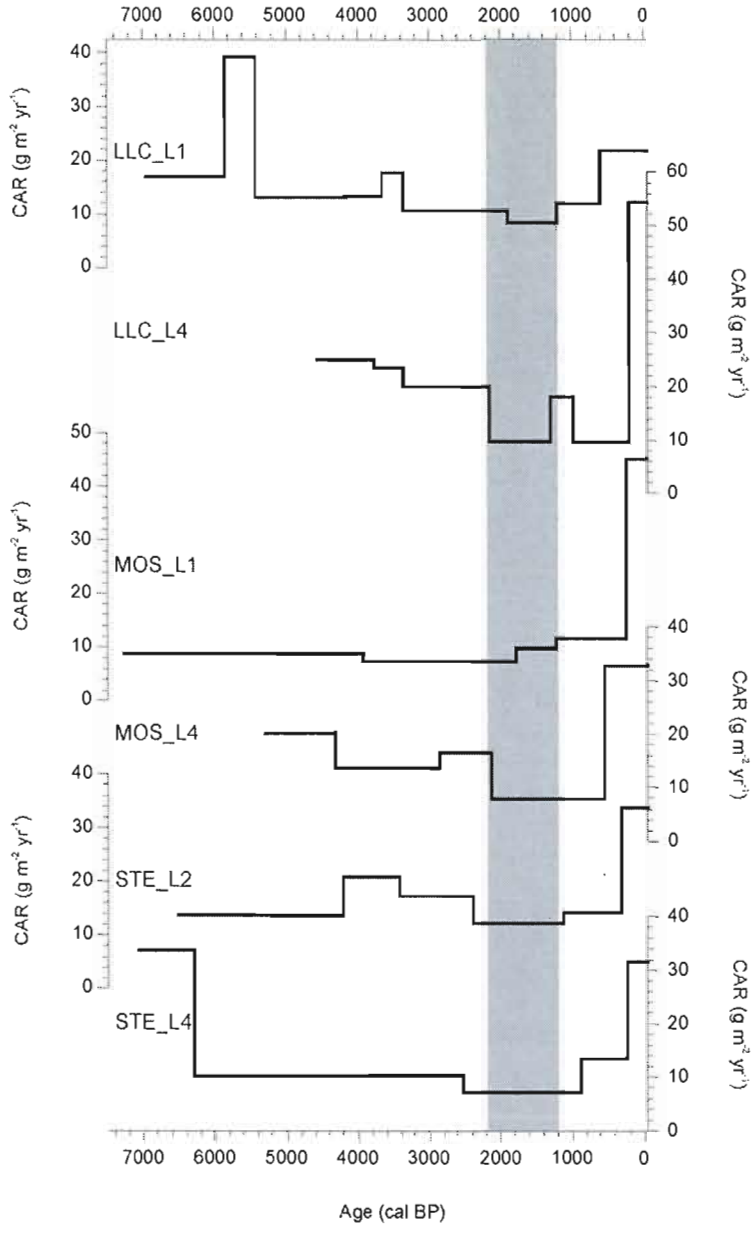

Figure 3.7. Long-term apparent carbon accumulation rates for each core. Grey zone indicates the regional late-Holocene period of low rates.

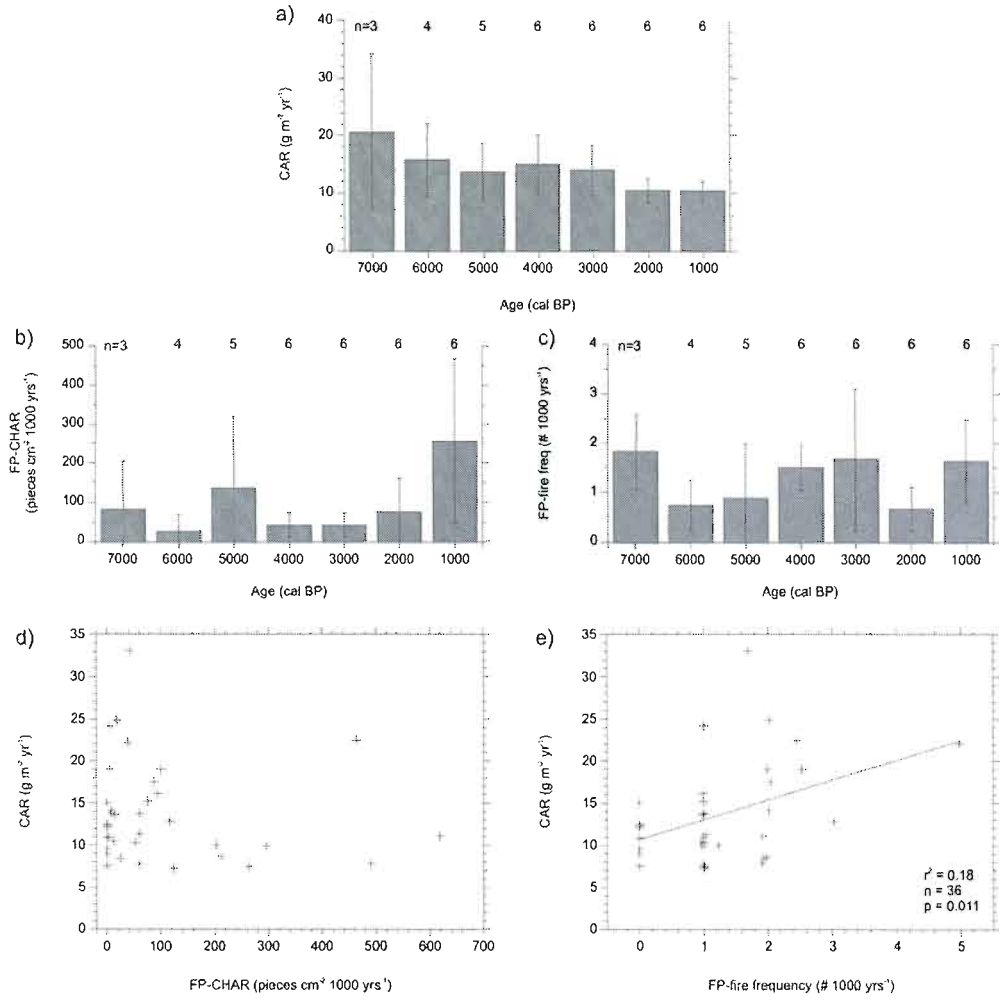

Figure 3.8. Carbon accumulation and fire regime reconstructions per 1000-yr time slice and regression results. (a) Mean carbon accumulation rates of all cores (± 2 SE). (b) Mean FP-CHAR and (c) FP-fire frequency of all cores (± 2 SE). (d) Regression of FP-CHAR and CAR and (e) FP-fire frequency and CAR.

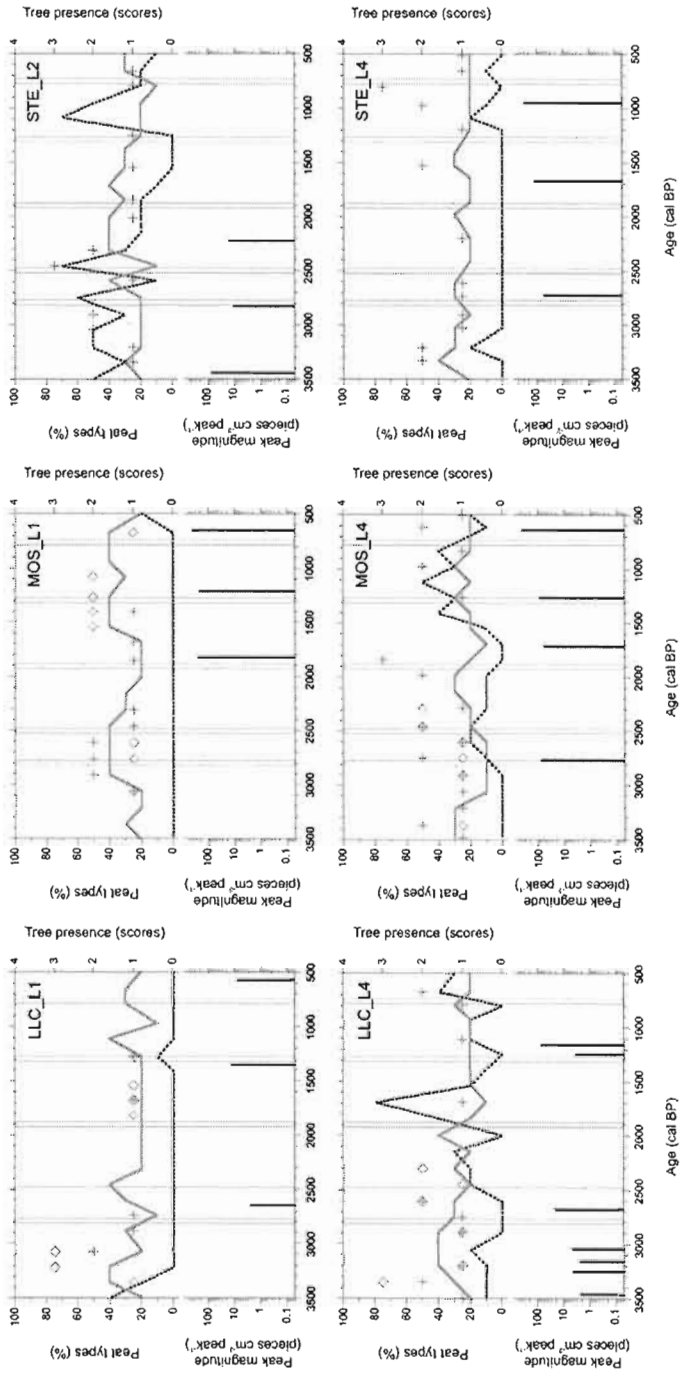

Figure 3.9. Vegetation and FP-fire peak magnitude records for the late-Holocene. Regional dry periods are indicated by grey vertical lines. Vegetation diagrams include *Sphagnum* (black dotted line), ligneous peat (grey line) and *Picea* (+) and *Larix* (◊) needle fragments.

Site	Core	Sample depth (cm)	UCIAMS lab. number	Material	^{14}C age (BP)	2σ range (cal BP)	Age (cal BP)
LLC	L1	45-46	58637	Sph stems	630±15	551-670	601
	L1	58-59	54956	Sph stems	1250±30	1078-1289	1207
	L1	68-69	58639	Sph stems, Larix/Picea leaf frs	1940±20	1810-1975	1889
	L1	99-100	57423	Sph stems	3125±15	3255-3390	3355
	L1	112-113	58638	Sph stems	3395±15	3586-3702	3656
	L1	130-131	54955	Sph stems	3780±25	4056-4269	4162
	L1	170-171	57418	Sph stems	4625±15	5295-5447	5418
	L1	210-211	64581	Sph stems	5035±20	5721-5903	5841
	L1	249-254	40365	Sph stems, Eric leaf frs	6055±20	6673-7218	6908
	L4	39-40	57417	Sph stems, Eric/Picea leaf frs	190±15	151-294	198
	L4	52-53	58632	Sph stems	1070±15	933-1075	978
	L4	61-63	58633	Sph stems, Eric leaf frs	1405±20	1216-1427	1301
	L4	78-79	57415	Sph stems, Larix/Eric leaf frs	2170±15	2129-2317	2152
	L4	127-128	57422	Sph stems	3135±15	3342-3435	3351
	L4	147-148	58635	Sph stems, Eric/Larix leaf frs	3495±20	3670-3826	3769
	L4	188-189	40368	Sph stems	4120±20	4520-4788	4586
MOS	L1	55-56	65378	Charcoal frs	190±15	147-379	243
	L1	77-78	65385	Charcoal frs	1260±20	1071-1308	1218
	L1	89-90	65389	Charcoal frs	1840±20	1678-2045	1787
	L1	117-119	65386	Charcoal, Cyp seeds	3625±20	3626-4312	3933
	L1	167-170	43474	Eric leaf frs; Cyp seeds	6420±20	6694-7657	7321
	L4	51-52	57425	Sph stems	455±15	502-550	545
	L4	73-74	58642	Sph stems	2165±20	2097-2285	2130
	L4	97-99	58641	Eric/Larix leaf frs	2750±20	2790-2911	2852
	L4	136-137	58640	Eric/Larix/Picea leaf frs	3835±15	4162-4326	4322
	L4	169-170	43476	Sph stems	4670±20	5318-5457	5323
STE	L2	44-45	67506	Sph stems	265±25	269-436	316
	L2	68-69	67507	Sph stems	1195±25	1031-1232	1123
	L2	98-99	67508	Sph stems	2380±25	2329-2588	2399
	L2	142-143	67509	Sph stems	3200±25	3360-3471	3415
	L2	180-181	67510	Sph stems	3820±25	4108-4363	4207
	L2	212-213	67511	Sph stems	4465±25	4977-5288	5182
	L2	244-246	40362	Sph stems	5760±20	6412-6725	6550
	L4	35-36	65384	Sph stems	135±20	70-287	224
	L4	48-49	67512	Charcoal frs	945±25	778-967	867
	L4	63-64	65380	Charcoal; Picea leaf frs	2455±20	2343-2708	2524
	L4	84-85	67513	Charcoal frs	3540±25	3703-3948	3826
	L4	125-126	65376	Charcoal, Picea/Eric leaf frs	5490±20	6189-6339	6288
	L4	174-176	40363	Sph stems; Larix leaf frs; Cyp seeds	6185±20	6976-7345	7090

Table 3.1. Radiocarbon dates selected by peatland and core.

Peatland	Core	Distance to forest (m)	Core length (cm)	Basal age (cal BP)	Median sample resolution (yr sample^{-1})	Median signal-to-noise index	F-fires (#)	FP-fires (#)	Mean FP-fire frequency (# 1000 yrs^{-1})	Carbon accumulation rate (g m^{-2} yr^{-1})	Mean peak magnitude FP-fires (pieces cm^{-2} peak^{-1})	
											Before 2000 cal BP	After 2000 cal BP
LLC	L1	26	252	6908	26	0.72	16	7	1.0	15.3	3.7	12.0
	L4	12	189	4586	22	0.80	18	10	2.2	19.4	9.1	39.0
MOS	L1	132	169	7321	47	0.91	15	7	0.9	10.3	118.3	293.7
	L4	39	170	5323	30	0.95	14	7	1.3	15.7	150.2	183.5
STE	L2	57	246	6550	24	0.84	19	8	1.2	15.9	31.5	9.8
	L4	34	176	7090	48	0.85	10	6	0.8	13.3	62.5	209.8

Table 3.2. Peatland core and fire regime characteristics.

CONCLUSION

The aims of the thesis may be summarized by 1) the quantification of total C stocks and rates of ecosystem C accumulation in three boreal peatlands and 2) the determination of the relative influence of varations in ecohydrological conditions and fire regimes within the Holocene hydroclimatic context. Data were obtained by field measurements and proxy analyses on multiple cores that covered spatial variability within each ecosystem. To accomplish the first objective, spatial extrapolation of peat C density and accumulation rates was performed. In addition, the use of plant macrofossil and testate amoeba analyses provided record of past vegetation assemblages and hydrological changes, which were matched to variations in reconstructed C accumulation rates. Finally, peatland fire history was obtained from stratigraphic macroscopic charcoal analyses, which was correlated with C accumulation rates.

1. Peatland C stocks

The mean of the C stock of each studied peatland is $203 \pm 28 \times 10^6$ kg ($\pm$ SE), with a mean area-weighted C mass of 91.6 kg m^{-2}. Considerable spatial variability in C sequestration patterns has been identified between peatlands. Firstly, the use of several cores collected in each peatland showed generally higher amounts of accumulated C in deep sections of the peatland than lateral cores. In accordance, ecosystem-scale C mass and C accumulation rates are generally lower than the quantifications from central cores (Fig. 1), with C mass overestimations attaining 23 to 61% of the mean ecosystem value.

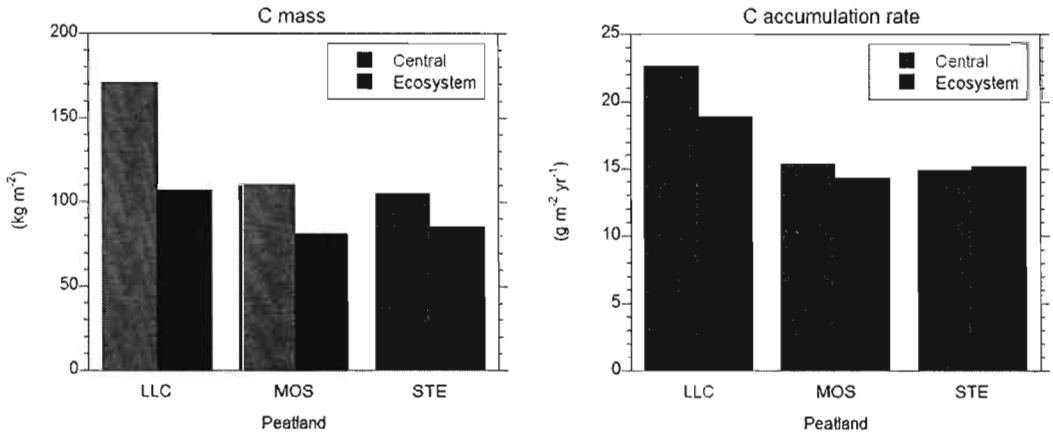

Figure 1. Comparisons of central core and ecosystem-scale reconstructions of C mass and C accumulation rate.

Comparisons between sites show that LLC bog as a whole has sequestered 242 × 10^6 kg C, whereas MOS and STE store 217 and 149 × 10^6 kg C, respectively, which is primarily the result of substantial differences in mean peat depth, probably induced by differences in local basin topography characteristics.

2. Ecohydrological changes

Reconstructions of vegetation and testate amoeba assemblages indicate that most sections of each peatland have accumulated for several millennia under ombrotrophic conditions. *Sphagnum* peat is a major contributor, especially in the central, deep sections of each peatland. Furthermore, *Sphagna* have been most stable in the core extracted from the thickest deposits of LLC bog. The central MOS bog record shows a more important presence of ligneous peat and less *Sphagna*, whereas peat composition in STE bog record is more dominated by ligneous and herbaceous peat. Testate amoeba assemblages are generally dominated by *Amphitrema flavum* and *Difflugia pulex*, with frequent episodic presence of *Amphitrema wrightianum*, *Hyalosphenia subflava* and the rotifer *Habrotrocha angusticollis*.

The transition from minerotrophic to ombrotrophic conditions occurred much earlier in the deep core from LLC (7210 cal BP) than in the deep cores from MOS (5450 cal BP) and

STE (4410 cal BP) bogs. Probably forced by persisting minerotrophic conditions, temporal variations in vegetation have been important in MOS and STE bog records, showing a general alternation of *Sphagnum* sections and Ericaceae and Cyperaceae assemblages at centennial scales. LLC bog record appears to have been more stable during the mid-Holocene, dominated by the presence of *Sphagnum* section *Acutifolia*. However, vegetation assemblages were highly variable in LLC bog between 3000 and 1000 cal BP, when increasing local presence of Ericaceae and Cyperaceae was reconstructed and testate amoeba-inferred water table depths show concomitant important fluctuations. These fluctuations match hydrological shifts reconstructed from MOS and STE records. Due to the regionally replicable trends, we may conclude that the last 3000 years have been characterized by hydroclimatic variations that caused fluctuating water table shifts and pressure on ecosystem stability, as demonstrated by collapsing *Sphagnum* cover in LLC bog and hydrological variations in all records. This interpretation is corroborated by previous late Holocene climate and environmental reconstructions in northern Quebec, that identified a general Neoglacial cooling (Arseneault and Sirois, 2004 ; Bhiry, Payette and Robert, 2007). Recurrent Little Ice Age (LIA) cooler and drier conditions during the last millennium might have resulted in years or decades of frozen horizons in the peatlands of the Eastmain region, causing an apparent net decrease in water table height. This hypothesis is supported by Thibault and Payette (2009) who demonstrated the contemporary persistence of residual permafrost in the La Grande Rivière area.

3. Eastmain peatland fire dynamics

Eastmain peatland fire regimes have been generally mild, with different peatland sections showing mean Holocene frequencies between 0.8 and 2.2 fires 1000 yrs^{-1}. Besides spatial variations, important temporal variations in charcoal peak magnitude and to a lesser extent fire frequency were found. Generally, important charcoal peaks and slightly higher fire frequencies were associated with the period between 2000 and 500 cal BP. As the past burning patterns show important variability at various spatial and temporal scales, our results suggest that fire regimes have likely been a function of the superposition of regional

(climatic), extra-local (e.g. landscape connectivity) and local (e.g. peatland microtopography) factors. As long-term (i.e. centennial-scale) shifts in vegetation appeared independent of fire events, generally more severe fires since 2000 cal BP may have been related to more important subdecadal drought, yet too short to induce a detectable shift in local vegetation. Reconstructions from boreal Quebec show intensifying upland fire regimes after 3000 cal BP, culminating between 2000 and 1000 cal BP (Arseneault and Sirois, 2004 ; Asselin and Payette, 2005 ; Carcaillet *et al.*, 2010), although some reconstructions show differential trends (e.g. Ali, Carcaillet and Bergeron, 2009). Discordant trends in upland and peatland fire regimes may be caused by differential sensitivity to climatic conditions. In peatlands, the influence of climate on hydrology is mediated by specific ecological feedback on climate conditions. Given the important variability in historical fire regimes, the use of multiple records from multiple peatlands is essential in future studies to accurately quantify regimes and identify driving factors.

4. Allogenic and autogenic influence on carbon accumulation: a question of scale

Comparing the deepest records from each peatland, Holocene C accumulation rates have been highest in the thickest, *Sphagnum*-dominated deposits of LLC bog at 22.6 g m^{-2} yr^{-1}. Mean Holocene C accumulation rates in MOS and STE are 15.4 and 14.9 g m^{-2} yr^{-1}, respectively. Although records have approximately the same age, the higher peat depth in LLC bog explains the larger area-weighted C mass (106.4 kg m^{-2}), compared to MOS (81.0 kg m^{-2}) and STE (85.8 kg m^{-2}). Taking into account spatial variability within the three peatlands, the regional area-weighted mean Holocene apparent rate of C accumulation attains 16.2 g m^{-2} yr^{-1}.

Reconstructions of long-term C sequestration show a general slowdown towards the late-Holocene, a trend that is observed in individual records (Fig. 2) as well as at the ecosystem level. As shown by peatland-specific age-depth models, vertical accumulation generally slowed down since the earliest stages of peatland development at 7500-7000 cal BP, while lateral expansion rates peaked roughly between 7000 and 6000 cal BP (Fig. 3). However, the combination of vertical accumulation and lateral expansion patterns

resulted in a regional culmination of C accumulation rates at the ecosystem level being delayed to 5000-3000 cal BP, thus representing an optimum in the combined effects of vertical and lateral accumulation.

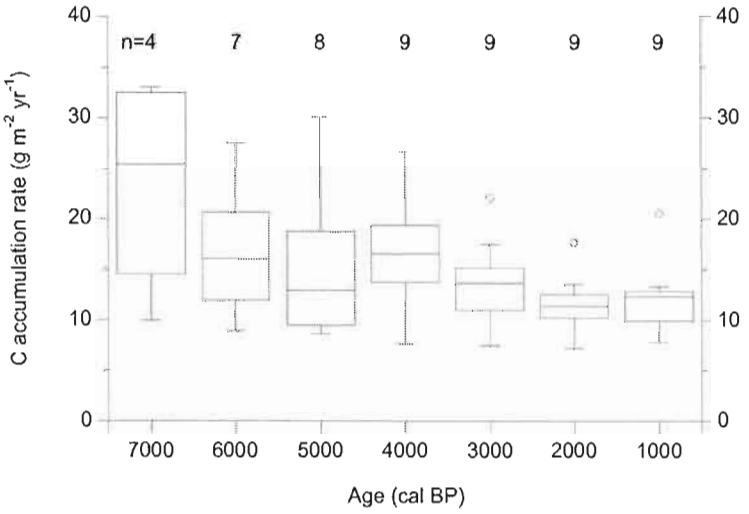

Figure 2. Box plot of Holocene C accumulation rates in 1000-yr bins of all central and lateral cores.

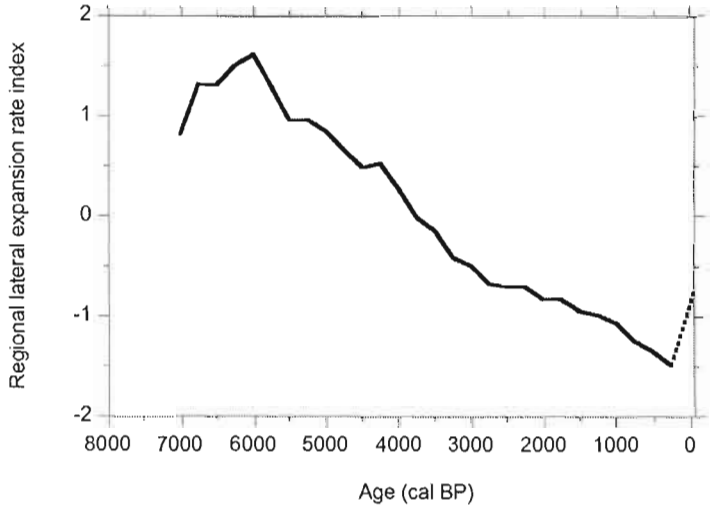

Figure 3. Synthesis of regional lateral expansion rates during the Holocene, expressed as index values.

We studied the influence of fire and vegetation effects on C accumulation rates by comparing vegetation and fire records to trends in C sequestration at the core scale. Episodes of rapid accumulation during the mid-Holocene in the deep core from LLC bog are associated with a stable cover of *Sphagnum* section *Acutifolia* and intermediate water table levels (10-15 cm below the surface). Slowdown of C accumulation has been associated with fluctuating water tables and changes in vegetation assemblages with more important Cyperaceae and ligneous species presence during the late-Holocene. Extremes in water table level are likely to have limiting effect on C accumulation. Deep water tables imply a long passage of litter in the acrotelm, and thus a high decay potential, whereas bryophytes associated with standing water or water tables close to the surface are sensitive to decomposition. However, high variability in C sequestration between cores indicates a complex response to driving factors. During the late-Holocene, generally colder conditions may have shortened the growing season length and diminished primary production rates (Mauquoy *et al.*, 2002). In addition, recurrent replicate events of low water tables, associated to the intrusion of Arctic air masses and cool/dry conditions, likely caused temporary increases in decomposition. These episodes thus had an additional limiting effect on regional peatland C sequestration.

Regression between fire regime traits and C accumulation rates showed that fires have not been the principal factor causing C accumulation rates to vary. Although fires have by definition a general limiting effect on C sequestration as C is released through combustion, this effect has been obscured on the long term by more dominant factors.

Despite an apparent link between hydroclimatic shifts and C accumulation rates, the identification of driving factors at the ecosystem level is much more complicated. Ecosystem-scale reconstructions of C sequestration patterns are more pertinent as they include the effects of lateral expansion, which is crucial to estimate global C dynamics during the Holocene (e.g. Korhola *et al.*, 2010). Ecosystem-scale reconstructions necessitate the use of multiple records that cover the spatial variability of the peatland basin. Extrapolation of accumulation rates obtained from these records to the entire ecosystem is relatively straightforward if rates from different peatland sections show uniform patterns, which was the case in LLC bog, and to a slightly lesser extent in MOS and STE bogs. Once a detailed image of past ecosystem rates of C sequestration has been created, however, the identification of factors controlling these rates appears much more complex. This complexity is primarily due to the fact that reconstructions

of local vegetation and water table heights identified in a core may not be as easily extrapolated to the scale of the ecosystem, as these factors show a much higher spatial variability than accumulation rates. Vegetation and hydrology in the Eastmain peatlands vary both at spatial scales of 1-10 m (e.g. following hummock-hollow patterning) and 100-1000 m (e.g. from open, wet central sections to relatively dry forest-peatland boundaries), whereas long-term accumulation rates vary only slightly over 100-1000 m (i.e. between central and lateral parts). Changes in lateral expansion have previously been linked to local variability in basin morphology, as long as climate is favourable for peat accumulation. The fact that core-scale vegetation history can not simply be extrapolated to the entire ecosystem largely inhibits the identification of the importance of ecohydrological change on peatland C sequestration at the ecosystem scale. Nevertheless, significant differences in mean C mass between peatlands, implies a strong local control (e.g. basin topography) on C sequestration.

Thus, although temporal variations in C accumulation rates may have been linked to changes in local ecohydrological conditions at decadal to centennial timescales, the relative influence of autogenic and allogenic (e.g. climate) factors at the ecosystem-scale remains difficult to quantify. However, significant differences in ecosystem characteristics indicate that Holocene C sequestration patterns have been influenced to a greater extent by local geomorphological conditions than what would be expected based on reconstructions from single cores.

5. Recommendations and future perspectives

The knowledge of past C accumulation patterns allows the identification of driving factors and the scale at which these factors operate. Once the importance of relevant factors is determined, accurate projections of future C accumulation may be possible. Single cores may actually represent C sequestration at the ecosystem scale only if the basin approximates a rectangular bucket and lateral expansion is absent. As lateral expansion is a generally important process in peatland dynamics and basins have variable slopes (Korhola, 1994 ; Mäkilä and Moisanen, 2007), single cores from peatlands are unable to accurately reflect sequestration at the ecosystem scale. Our results show that continuing lateral expansion may

cause an increasing trend in peat accumulation rates at the scale of the ecosystem even when vertical accumulation decelerates. The importance of expansion potential on peatland dynamics has previously been noted with respect to microform stability (Belyea and Clymo, 2001). Thus, we suggest that expansion patterns should be taken into account considering both reconstructions of C sequestration and interpretation of past records of ecohydrological change. Using a novel approach to quantify C sequestration at the ecosystem scale, we provide a new piece to the global image of terrestrial C stocks by ecosystem-scale quantifications and reconstructions of accumulation patterns from a largely unexplored region.

This thesis reinforces previous climatic rcconstructions from northern Quebec using testate-amoeba-inferred water table fluctuations, a tool rarely employed in this part of the North American continent. In order to increase the comprehension on the influence of short-term shifts in ecohydrological conditions on C accumulation rates, e.g. shifts associated with the LIA, dating, testate amoeba and plant macrofossil analyses should be performed at high (e.g. subcentennial to decadal) resolution.

Finally, the reconstruction of unforested peatland fire regimes under specific climatic conditions improves the comprehension on peatland burning potential through a better understanding of scales and factors relevant to fire events. The use of multiple records appears essential in the reconstruction of open peatland fire regimes, as spatial variability in burning patterns is high. We show that open peatland burning events are highly linked to short-scale and local factors, yet peatland fire regimes are initially dependent on the occurrence of upland fire. Increasing trends in peatland fire regimes were reconstructed after 2200 cal BP, while upland regimes decreased. More severe late-Holocene peatland fires may have been associated to important dry shifts in peatlands, linked to cold and dry climatic conditions. Decreasing upland forest fire potential may have been the result of a differential sensitivity to climatic conditions. Thus, in order to accurately project future peatland fire scenarios, one should take account of this differential sensitivity and thus open peatlands should be modelled separately. Open peatlands thus differ markedly from boreal forested peatlands in western Canada, where peatlands have been reported to burn as often as uplands, importantly influencing long-term C sequestration (Kuhry, 1994 ; Turetsky *et al.*, 2004).

Climatic differences between the drier western and more humid eastern Canada support these variations in peatland dynamics.

Twenty-first century C sequestration dynamics in the Eastmain peatlands may be relatively independent of autogenic processes, as the latter generally act at millennial timescales (Charman *et al.*, 2006). Given the intensity of 21st century climate projections, trends in C sequestration over this period are thus likely to be strongly climate-related. Climate projections for 2040-2060 relative to 1970-1990, generated by the Canadian Regional Climate Model (A2 scenario) for the Eastmain region, indicate increases in summer and winter mean temperatures of 2.0-2.5°C and 3.0-4.0°C and ~10% and ~15% increases in summer and winter precipitation, respectively (Plummer *et al.*, 2006). These changes may affect the regional peatland water balance, yet the impact of these changes probably depends on the relative effects on precipitation and evapotranspiration. General warming will cause an increasing potential for *Sphagnum* growth, although a concomitant net decrease in water table levels may limit bryophyte productivity while increasing decay. Projections of increases in both winter temperature and precipitation may affect peatland hydrology in a complex way. Due to increasing winter temperatures, the duration of the snowfall period is likely to diminish, yet increasing precipitation might partly compensate for this effect and thus snow depths may be maintained. Peatland water table fluctuations may be generally driven by summer precipitation and to a lesser extent summer temperature (Booth, 2010 ; Charman *et al.*, 2009), but winter precipitation and temperature may be additionally important in specific settings (e.g. Lamentowicz *et al.*, 2010). Additional research on driving factors for boreal peatlands, including peatlands with important snow cover and (relict) permafrost dynamics is necessary to better project the effects of future climate on hydrology, and thus C sequestration.

As fire regimes were shown to be relatively independent of local vegetation, we hypothesize that decadal-scale decreases in water table may be the best predictors for fire occurrence. Although future warming may imply decreasing water table levels, past events of dry peatland surfaces were rather linked to dry, cold conditions. The last 150 years have shown general increases in temperature and precipitation, with a trend of higher water tables and reduced drought frequency in northern Quebec peatlands and uplands (Arlen-Pouliot and Bhiry, 2005 ; Bergeron and Archambault, 1993 ; Lesieur, Gauthier and Bergeron, 2002 ;

Loisel and Garneau, 2010 ; Payette and Delwaide, 2004 ; Payette *et al.*, 2004). Nevertheless, these recent increases in peatland water tables may be primarily explained by melting of LIA permafrost rather than increases in precipitation. As no relict permafrost is present in the Eastmain region, future climate may well cause decreasing water tables and increasing the potential for fire occurrence, possibly decreasing C sequestration potential.

REFERENCES

Aaby, B. 1976. «Cyclic climatic variations in climate over the past 5,500 yr reflected in raised bogs». *Nature*. vol. 263, no 5575, p. 281-284.

Ali, A. A., C. Carcaillet and Y. Bergeron. 2009. «Long-term fire frequency variability in the eastern Canadian boreal forest: the influences of climate vs. local factors». *Global Change Biology*. vol. 15, no 5, p. 1230-1241.

Arlen-Pouliot, Y., and N. Bhiry. 2005. «Palaeoecology of a palsa and a filled thermokarst pond in a permafrost peatland, subarctic Québec, Canada». *The Holocene*. vol. 15, no 3, p. 408-419.

Arseneault, D., and L. Sirois. 2004. «The millennial dynamics of a boreal forest stand from buried trees». *Journal of Ecology*. vol. 92, p. 490-504.

Asselin, H., and S. Payette. 2005. «Late Holocene opening of the forest tundra landscape in northern Québec, Canada». *Global Ecology and Biogeography*. vol. 14, no 4, p. 307-313.

Barber, K. E., F. M. Chambers, D. Maddy, R. Stoneman and J. S. Brew. 1994. «A sensitive high-resolution record of late Holocene climatic change from a raised bog in northern England». *The Holocene*. vol. 4, no 2, p. 198-205.

Beaulieu-Audy, V., M. Garneau, P. J. H. Richard and H. Asnong. 2009. «Holocene palaeoecological reconstruction of three boreal peatlands in the La Grande Riviere region, Quebec, Canada». *The Holocene*. vol. 19, no 3, p. 459-476.

Beilman, D. W., D. H. Vitt, J. S. Bhatti and S. Forest. 2008. «Peat carbon stocks in the southern Mackenzie River Basin: uncertainties revealed in a high-resolution case study». *Global Change Biology*. vol. 14, no 6, p. 1221-1232.

Beilman, D. W., G. M. MacDonald, L. C. Smith and P. J. Reimer. 2009. «Carbon accumulation in peatlands of West Siberia over the last 2000 years». *Global Biogeochemical Cycles*. vol. 23. GB1012, doi:10.1029/2007GB003112

Belyea, L. R. 1996. «Separating the effects of litter quality and microenvironment on decomposition rates in a patterned peatland». *Oikos*. vol. 77, no 3, p. 529-539.

Belyea, L. R., and R. S. Clymo. 2001. «Feedback control on the rate of peat formation». *Proceedings: Biological Sciences*. vol. 268, no 1473, p. 1315-1321.

Belyea, L. R., and N. Malmer. 2004. «Carbon sequestration in peatland: patterns and mechanisms of response to climate change». *Global Change Biology*. vol. 10, no 7, p. 1043-1052.

Bergeron, Y., and S. Archambault. 1993. «Decreasing frequency of forest fires in the southern boreal zone of Quebec and its relation to global warming since the end of the 'Little Ice Age'». *The Holocene*. vol. 3, no 3, p. 255-259.

Bhiry, N., M. Garneau and L. Filion. 2000. «Macrofossil record of a middle Holocene drop in relative sea level at the St. Lawrence Estuary, Quebec». *Quaternary Research*. vol. 54, no 2, p. 228-237.

Bhiry, N., S. Payette and É. C. Robert. 2007. «Peatland development at the arctic tree line (Québec, Canada) influenced by flooding and permafrost». *Quaternary Research*. vol. 67, no 3, p. 426-437.

Booth, R. K. 2008. «Testate amoebae as proxies for mean annual water-table depth in Sphagnum-dominated peatlands of North America». *Journal of Quaternary Science*. vol. 23, no 1, p. 43-57.

Booth, R. K. 2010. «Testing the climate sensitivity of peat-based paleoclimate reconstructions in mid-continental North America». *Quaternary Science Reviews*. vol. 29, no 5-6, p. 720-731.

Borren, W., W. Bleuten and E. D. Lapshina. 2004. «Holocene peat and carbon accumulation rates in the southern taiga of western Siberia». *Quaternary Research*. vol. 61, no 1, p. 42-51.

Bubier, J. L., T. R. Moore and S. Juggins. 1995. «Predicting methane emission from bryophyte distribution in northern Canadian peatlands». *Ecology*. vol. 76, no 3, p. 677-693.

Carcaillet, C., Y. Bergeron, P. J. H. Richard, B. Fréchette, S. Gauthier and Y. T. Prairie. 2001. «Change of fire frequency in the eastern Canadian boreal forests during the Holocene: does vegetation composition or climate trigger the fire regime?». *Journal of Ecology*. vol. 89, no 6, p. 930-946.

Carcaillet, C., P. J. H. Richard, Y. Bergeron, B. Fréchette and A. A. Ali. 2010. «Resilience of the boreal forest in response to Holocene fire-frequency changes assessed by pollen diversity and population dynamics». *International Journal of Wildland Fire*. vol. 19, p. 1026-1039.

Chambers, F. M., and D. J. Charman. 2004. «Holocene environmental change: contributions from the peatland archive». *The Holocene*. vol. 14, no 1, p. 1-6.

Charman, D. J. 2002. *Peatlands and environmental change*. Chichester: John Wiley & Sons Ltd., 301 p.

Charman, D. J. 2007. «Summer water deficit variability controls on peatland water-table changes: implications for Holocene palaeoclimate reconstructions». *The Holocene*. vol. 17, no 2, p. 217-227.

Charman, D. J., K. E. Barber, M. Blaauw, P. G. Langdon, D. Mauquoy, T. J. Daley, P. D. M. Hughes and E. Karofeld. 2009. «Climate drivers for peatland palaeoclimate records». *Quaternary Science Reviews*. vol. 28, no 19-20, p. 1811-1819.

Charman, D. J., D. W. Beilman, M. Blaauw, R. K. Booth, S. Brewer, F. M. Chambers, J. A. Christen, A. Gallego-Sala, S. P. Harrison, P. D. M. Hughes, S. T. Jackson, A. Korhola, D. Mauquoy, F. J. G. Mitchell, I. C. Prentice, M. van der Linden, F. de Vleeschouwer, Z. C. Yu, I. E. Bauer, M. Garneau, V. Hohl, E. Karofeld, G. Le Roux, R. Moschen, J. E. Nichols, N. R. Phadtare, N. Rausch, G. T. Swindles, M. Väliranta, S. van Bellen, B. van Geel and D. H. Vitt. submitted. «Carbon-cycle implications of climate-driven changes in peat accumulation during the last millennium».

Charman, D. J., A. Blundell and ACCROTELM members. 2007. «A new European testate amoebae transfer function for palaeohydrological reconstruction on ombrotrophic peatlands». *Journal of Quaternary Science*. vol. 22, no 3, p. 209-221.

Clymo, R. S. 1983. «Peat». In *Mires: swamp, bog, fen and moor*, 4A, A. J. P. Gore, p. 159-224. Amsterdam: Elsevier.

Clymo, R. S. 1984. «The limits to peat bog growth». *Philosophical Transactions of the Royal Society of London. Series B, Biological Sciences*. vol. 303, no 1117, p. 605-654.

Clymo, R. S., J. Turunen and K. Tolonen. 1998. «Carbon accumulation in peatland». *Oikos*. vol. 81, no 2, p. 368-388.

Dupont, L. M. 1986. «Temperature and rainfall variation in the Holocene based on comparative palaeoecology and isotope geology of a hummock and a hollow (Bourtangerveen, The Netherlands)». *Review of Palaeobotany and Palynology*. vol. 48, no 1-3, p. 71-88.

Eppinga, M. B., M. Rietkerk, W. Borren, E. D. Lapshina, W. Bleuten and M. J. Wassen. 2008. «Regular surface patterning of peatlands: Confronting theory with field data». *Ecosystems*. vol. 11, no 4, p. 520-536.

Filion, L. 1984. «A relationship between dunes, fire and climate recorded in the Holocene deposits of Quebec». *Nature*. vol. 309, no 5968, p. 543-546.

Frolking, S., and N. T. Roulet. 2007. «Holocene radiative forcing impact of northern peatland carbon accumulation and methane emissions». *Global Change Biology*. vol. 13, no 5, p. 1079-1088.

Glaser, P. H., B. Hansen, C. S., D. I. Siegel, A. S. Reeve and P. J. Morin. 2004. «Rates, pathways and drivers for peatland development in the Hudson Bay Lowlands, northern Ontario, Canada». *Journal of Ecology*. vol. 92, no 6, p. 1036-1053.

Gorham, E. 1991. «Northern peatlands: role in the carbon cycle and probable responses to climatic warming». *Ecological Applications*. vol. 1, no 2, p. 182-195.

Gorham, E., J. A. Janssens and P. H. Glaser. 2003. «Rates of peat accumulation during the postglacial period in 32 sites from Alaska to Newfoundland, with special emphasis on northern Minnesota». *Canadian Journal of Botany*. vol. 81, no 5, p. 429-438.

Hellberg, E., M. Niklasson and A. Granström. 2004. «Influence of landscape structure on patterns of forest fires in boreal forest landscapes in Sweden». *Canadian Journal of Forest Research*. vol. 34, p. 332-338.

Hughes, P. D. M. 2000. «A reappraisal of the mechanisms leading to ombrotrophy in British raised mires». *Ecology Letters*. vol. 3, no 1, p. 7-9.

Hughes, P. D. M., and L. Dumayne-Peaty. 2002. «Testing theories of mire development using multiple successions at Crymlyn Bog, West Glamorgan, South Wales, UK». *Journal of Ecology*. vol. 90, no 3, p. 456-471.

IPCC. 2007. Climate change 2007: The physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. S. Solomon, D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Averyt, M. Tignor et H. L. Miller. Cambridge: 996 p.

Kaufman, D. S., T. A. Ager, N. J. Anderson, P. M. Anderson, J. T. Andrews, P. J. Bartlein, L. B. Brubaker, L. L. Coats, L. C. Cwynar, M. L. Duvall, A. S. Dyke, M. E. Edwards, W. R. Eisner, K. Gajewski, A. Geirsdóttir, F. S. Hu, A. E. Jennings, M. R. Kaplan, M. W. Kerwin, A. V. Lozhkin, G. M. MacDonald, G. H. Miller, C. J. Mock, W. W. Oswald, B. L. Otto-Bliesner, D. F. Porinchu, K. Rühland, J. P. Smol, E. J. Steig and B. B. Wolfe. 2004. «Holocene thermal maximum in the western Arctic (0-180°W)». *Quaternary Science Reviews*. vol. 23, no 5-6, p. 529-560.

Kerwin, M. W., J. T. Overpeck, R. S. Webb and K. H. Anderson. 2004. «Pollen-based summer temperature reconstructions for the eastern Canadian boreal forest, subarctic, and Arctic». *Quaternary Science Reviews*. vol. 23, no 18-19, p. 1901-1924.

Korhola, A. 1994. «Radiocarbon evidence for rates of lateral expansion in raised mires in southern Finland». *Quaternary Research*. vol. 42, no 3, p. 299-307.

Korhola, A., M. Ruppel, H. Seppä, M. Väliranta, T. Virtanen and J. Weckström. 2010. «The importance of northern peatland expansion to the late-Holocene rise of atmospheric methane». *Quaternary Science Reviews*. vol. 29, no 5-6, p. 611-617.

Kuhry, P. 1994. «The role of fire in the development of Sphagnum-dominated peatlands in western boreal Canada». *Journal of Ecology*. vol. 82, no 4, p. 899-910.

Kuhry, P., and D. H. Vitt. 1996. «Fossil carbon/nitrogen ratios as a measure of peat decomposition». *Ecology*. vol. 77, no 1, p. 271-275.

Lamentowicz, M., W. van der Knaap, L. Lamentowicz, J. F. N. van Leeuwen, E. A. D. Mitchell, T. Goslar and C. Kamenik. 2010. «A near-annual palaeohydrological study based on testate amoebae from a sub-alpine mire: Surface wetness and the role of climate during the instrumental period». *Journal of Quaternary Science*. vol. 25, no 2, p. 190-202.

Lesieur, D., S. Gauthier and Y. Bergeron. 2002. «Fire frequency and vegetation dynamics for the south-central boreal forest of Quebec, Canada». *Canadian Journal of Forest Research*. vol. 32, no 11, p. 1996-2009.

Loisel, J., and M. Garneau. 2010. «Late Holocene paleoecohydrology and carbon accumulation estimates from two boreal peat bogs in eastern Canada: Potential and limits of multi-proxy archives». *Palaeogeography, Palaeoclimatology, Palaeoecology*. vol. 291, no 3-4, p. 493-533.

Lynch, J. A., J. S. Clark and B. J. Stocks. 2004. «Charcoal production, dispersal, and deposition from the Fort Providence experimental fire: Interpreting fire regimes from charcoal records in boreal forests». *Canadian Journal of Forest Research*. vol. 34, no 8, p. 1642-1656.

MacDonald, G. M., D. W. Beilman, K. V. Kremenetski, Y. Sheng, L. C. Smith and A. A. Velichko. 2006. «Rapid early development of circumarctic peatlands and atmospheric CH_4 and CO_2 variations». *Science*. vol. 314, no 5797, p. 285-288.

Mäkilä, M. 1997. «Holocene lateral expansion, peat growth and carbon accumulation on Haukkasuo, a raised bog in southeastern Finland». *Boreas*. vol. 26, no 1, p. 1-14.

Mäkilä, M., and M. Moisanen. 2007. «Holocene lateral expansion and carbon accumulation of Luovuoma, a northern fen in finnish Lapland». *Boreas*. vol. 36, no 2, p. 198-210.

Mansuy, N., S. Gauthier, A. Robitaille and Y. Bergeron. 2010. «The effects of surficial deposit-drainage combinations on spatial variations of fire cycles in the boreal forest of eastern Canada.». *International Journal of Wildland Fire*. vol. 19, no 8, p. 1083-1098.

Mauquoy, D., T. Engelkes, M. H. M. Groot, F. Markesteijn, M. G. Oudejans, J. van der Plicht and B. van Geel. 2002. «High-resolution records of late-Holocene climate change and carbon accumulation in two north-west European ombrotrophic peat bogs». *Palaeogeography, Palaeoclimatology, Palaeoecology*. vol. 186, no 3-4, p. 275-310.

Payette, S., and R. Gagnon. 1985. «Late Holocene deforestation and tree regeneration in the forest-tundra of Quebec». *Nature*. vol. 313, no 6003, p. 570-572.

Payette, S., and A. Delwaide. 2004. «Dynamics of subarctic wetland forests over the past 1500 years ». *Ecological Monographs*. vol. 74, no 3, p. 373-391.

Payette, S., and L. Rochefort. 2001. *Écologie des tourbières du Québec-Labrador* Trad. de: *French*. Ste-Foy: Les presses de l'Université Laval, 621 p.

Payette, S., L. Gauthier and I. Grenier. 1986. «Dating ice-wedge growth in subarctic peatlands following deforestation». *Nature*. vol. 322, p. 724-727.

Payette, S., A. Delwaide, M. Caccianiga and M. Beauchemin. 2004. «Accelerated thawing of subarctic peatland permafrost over the last 50 years». *Geophysical Research Letters*. vol. 31, no 18, p. L18208.

Payette, S., C. Morneau, L. Sirois and M. Desponts. 1989. «Recent fire history of the northern Quebec biomes». *Ecology*. vol. 70, no 3, p. 656-673.

Pitkänen, A., J. Turunen and K. Tolonen. 1999. «The role of fire in the carbon dynamics of a mire, eastern Finland». *The Holocene*. vol. 9, no 4, p. 453-462.

Plummer, D. A., D. Caya, A. Frigon, H. Côté, M. Giguère, D. Paquin, S. Biner, R. Harvey and R. de Elia. 2006. «Climate and climate change over North America as simulated by the Canadian RCM». *Journal of Climate*. vol. 19, no 13, p. 3112-3132.

Thibault, S., and S. Payette. 2009. «Recent permafrost degradation in bogs of the James Bay area, northern Quebec, Canada». *Permafrost and Periglacial Processes*. vol. 20, no 4, p. 383-389.

Thormann, M. N., and S. E. Bayley. 1997. «Aboveground plant production and nutrient content of the vegetation in six peatlands in Alberta, Canada». *Plant Ecology*. vol. 131, no 1, p. 1-16.

Turetsky, M., K. Wieder, L. Halsey and D. Vitt. 2002. «Current disturbance and the diminishing peatland carbon sink». *Geophysical Research Letters*. vol. 29, no 11, 1526, p. doi:10.1029/2001GL014000.

Turetsky, M. R., B. D. Amiro, E. Bosch and J. S. Bhatti. 2004. «Historical burn area in western Canadian peatlands and its relationship to fire weather indices». *Global Biogeochemical Cycles*. vol. 18, no 4, p. GB4014.

Turunen, J., T. Tahvanainen, K. Tolonen and A. Pitkänen. 2001. «Carbon accumulation in West Siberian mires, Russia». *Global Biogeochemical Cycles*. vol. 15, no 2, p. 285-296.

Turunen, J., N. T. Roulet, T. R. Moore and P. J. H. Richard. 2004. «Nitrogen deposition and increased carbon accumulation in ombrotrophic peatlands in eastern Canada». *Global Biogeochemical Cycles*. vol. 18, no GB3002, p. doi:10.1029/2003GB002154

van Breemen, N. 1995. «How Sphagnum bogs down other plants». *Trends in Ecology and Evolution*. vol. 10, no 7, p. 270-275.

van Geel, B., and W. G. Mook. 1989. «High-resolution ^{14}C dating of organic deposits using natural atmospheric ^{14}C variations». *Radiocarbon*. vol. 31, no 2, p. 151-155.

Vardy, S. R., B. G. Warner, J. Turunen and R. Aravena. 2000. «Carbon accumulation in permafrost peatlands in the Northwest Territories and Nunavut, Canada». *The Holocene*. vol. 10, no 2, p. 273-280.

Viau, A. E., K. Gajewski, M. C. Sawada and P. Fines. 2006. «Millennial-scale temperature variations in North America during the Holocene». *Journal of Geophysical Research*. vol. 111.

Vitt, D. H., L. A. Halsey, I. E. Bauer and C. Campbell. 2000. «Spatial and temporal trends in carbon storage of peatlands of continental western Canada through the Holocene». *Canadian Journal of Earth Sciences*. vol. 37, no 5, p. 683-693.

Waddington, J. M., and N. T. Roulet. 1999. «Carbon balance of a boreal patterned peatland». *Global Change Biology*. vol. 6, p. 87-97.

Warner, B. G., and D. J. Charman. 1994. «Holocene changes on a peatland in northwestern Ontario interpreted from testate amoebae (Protozoa) analysis». *Boreas*. vol. 23, no 3, p. 270-279.

Wieder, R. K., K. D. Scott, K. Kamminga, M. A. Vile, D. H. Vitt, T. Bone, B. Xu, B. W. Benscoter and J. S. Bhatti. 2009. «Postfire carbon balance in boreal bogs of Alberta, Canada». *Global Change Biology*. vol. 15, no 1, p. 63-81.

Woodland, W. A., D. J. Charman and P. C. Sims. 1998. «Quantitative estimates of water tables and soil moisture in Holocene peatlands from testate amoebae». *The Holocene*. vol. 8, no 3, p. 261-273.

Yu, Z., D. H. Vitt, I. D. Campbell and M. J. Apps. 2003. «Understanding Holocene peat accumulation pattern of continental fens in western Canada». *Canadian Journal of Botany*. vol 81, p. 267-282

Yu, Z. 2006. «Holocene carbon accumulation of fen peatlands in boreal western Canada: a complex ecosystem response to climate variation and disturbance». *Ecosystems*. vol. 9, no 8, p. 1278-1288.

Yu, Z., D. W. Beilman and M. C. Jones. 2009. «Sensitivity of northern peatland carbon dynamics to Holocene climate change». In *Carbon cycling in northern peatlands,*

Andrew J. Baird, Lisa R. Belyea, Xavier Comas, A. S. Reeve et Lee D. Slater, p. 55-69. Washington: American Geophysical Union.

Yu, Z., J. Loisel, D. P. Brosseau, D. W. Beilman and S. J. Hunt. 2010. «Global peatland dynamics since the Last Glacial Maximum». *Geophysical Research Letters*. vol. 37, no 13, p. L13402.

Yu, Z., D. H. Vitt, I. D. Campbell and M. J. Apps. 2003. «Understanding Holocene peat accumulation pattern of continental fens in western Canada». *Canadian Journal of Botany*. vol. 81, p. 267-282.

Zoltai, S. C., L. A. Morrissey, G. P. Livingston and W. J. de Groot. 1998. «Effects of fires on carbon cycling in North American boreal peatlands». *Environmental Reviews*. vol. 6, no 1, p. 13-24.